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The palaeoenvironment and palaeoecology of a middle Jurassic vertebrate-bearing fen-type paleosol in a coastal carbonate regime.

The Palaeoenvironment and Palaeoecology of a Middle Jurassic Vertebrate-bearing Fen-type Paleosol in a Coastal Carbonate Regime.

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Submitted in fulfilment of the requirements for the degree of

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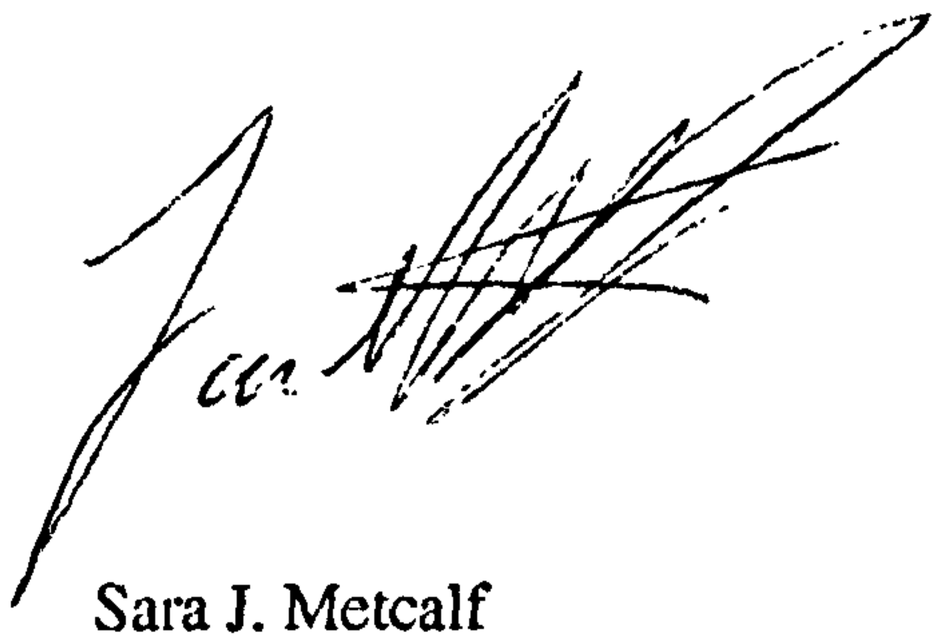
CONTAINS

PULLOUTS

This thesis and the conclusions herein are the results of my own original research except where due references have been given

Signed



Sara J. Metcalf

12-6-1995

Abstract

A palaeoenvironmental model for a regional marine regression in the Bathonian (Middle Jurassic) Great Oolite Group of the British Midlands is provided. The reconstruction is based upon a detailed analysis of a freshwater fossiliferous clay unit, interpreted as a fen-type paleosol profile developing within a karstic hollow or 'doline' in the predominantly marine oolitic limestones of the Chipping Norton Formation upon the exposed Cotswolds-Weald carbonate shelf. The implications of the model suggest that regressive-marsh/coastal-plain facies were much more widespread in the early Bathonian than previously considered and a tectonic control upon sedimentation is proposed, associated with rifting in the North Atlantic.

The Middle Jurassic of central and south-west England has produced several rich terrestrial vertebrate sites. The most recently excavated Hornsleasow Quarry, Gloucestershire, is the oldest known British Bathonian locality (*zigzag* zone), and has already produced a diverse fauna of macro- and microvertebrate remains. Both aquatic and terrestrial biotas are represented, and there is some suggestion of a marine influence within this predominantly fresh-water to brackish, boggy environment. The fossil assemblage is also crucial in enabling reconstruction of a Middle Jurassic terrestrial tetrapod palaeocommunity and providing information upon palaeoclimatic conditions prevalent at the time of deposition.

The macrovertebrate remains are weathered and heavily mineralised. They largely represent the partial remains of a disarticulated individual of the sauropod dinosaur *Cetiosaurus* sp., and are confined to the lowest unit of the lens, directly upon the palaeokarstic surface. Their poor preservation suggests slight sub-aerial exposure and possible transport prior to burial within the paleosol. The abundance of poorly aligned carbonised wood debris, mixed with the cetiosaur material, suggests only superficial palaeocurrents within this lower unit.

Microvertebrate remains occur throughout the entire clay lens, and are of a disassociated and fragmental nature. Skeletal elements are dominated by isolated teeth, jaw elements, vertebrae, broken limb elements, and podia, as well as fragmental fish scales and archosaur dermal armour. The material is much less degraded, and breakages have a much fresher appearence than in the larger material. The microvertebrates are interpreted as having suffered little transport, and are probably indigenous. Coprolites, wood debris, and scattered invertebrate remains (fresh-water gastropods, ostracods, and some marine foraminiferans) also occur. The presence in the clay lens of much open marine invertebrate debris in various states of preservation is explained by the *in situ* karstic, and soil-forming, weathering of the surrounding oolites.

This thesis is dedicated to the memory of my grandpa, Ronald Wright, who bought my first dinosaur book and who never faltered in his dream that one day this precocious child might become a vertebrate palaeontologist.

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I have been extremely fortunate to have worked in collaboration with and received much insight upon all aspects of the research from friends and colleagues including Drs. Jim Cole, Susan Evans, Simon Haslett, Tim Jones, Amanda Kear, Andrew Milner, Mike Simms, Mike Taylor, Charlie Underwood, and David Unwin. Liz Cook and Suzanne Ashford also provided me with useful ideas and perspectives on the nature of taphonomy. Countless friends and colleagues have accompanied me on fieldwork into the wilder parts of the Cotswolds. Fieldwork upon the Isles of Skye and Eigg, was supported by the Bryan Patterson Award from the Society of Vertebrate Paleontology and assistance in the field provided by Mr. Martin Gagg (thanks for going up that cliff for me, Martin). Mr. Andrew Ross was nice and clever enough to help me with countless unfathomable statistical tests and gave beneficial massages. Finally, I'd like to thank my family for having the courage and confidence to support me throughout my four-year limbo in academia. Thanks for understanding, Dad.

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PART C : Palaeontology, palaeoecology and taphonomy of the Hornsleasow biota

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Chapter One Introduction

1.1. Introduction

The Jurassic System is named from the succession of rocks of that age in the Jura Mountains of eastern France and Switzerland. The sequence in this region was first recognised as separate from the underlying Triassic by von Humboldt (1795), and named by him the 'Calcaires de Jura'. Rocks of this age were first recognised in the British Isles by William Smith, between 1797-1817 in southern England, and these were grouped together as the 'Oolite Formation' (Buckland, 1818) or the 'Oolite Series' (Conybeare & Phillips, 1822). In 1829, Brongniart devised the name 'Terrains Jurassique' for the succession, but restricted the term to mean only the 'Lower Oolites' of Conybeare & Phillips (1822). This 'Jurassique' terminology coexisted in Great Britain well into the latter half of 19th century with the earlier names : Lias and Oolites. However, the term 'Jurassic' received wider recognition in British geology after Wright (1872) produced a paper upon the correlation of the British and French oolitic limestones.

Oppel (1856) devised a scheme of biostratigraphic zonation for the Jurassic period, based upon ammonite fossils, and grouped his zones into various stages and these subsequently into three major divisions. This scheme is still in use at the present, and the three major epochs are : the Lower Jurassic or 'Lias'; the Middle Jurassic or 'Dogger' (corresponding to the Lower Oolites of Coneybeare & Phillips, 1822); and the Upper Jurassic or 'Malm' (which corresponds to Coneybeare & Phillips' Middle and Upper Oolites (1822)).

The Jurassic System in the British Isles has been very important in the field of stratigraphy as the "proving ground for many major advances in stratigraphical practice" (H.S. Torrens in Cope *et al.*, 1980a, p.2). This is largely because of the extreme abundance of fossils in these rocks, and in particular ammonites, on which the biostratigraphic subdivision of the period is based.

Rocks of Jurassic age occur in the British Isles in a long, sinuous outcrop, running from Dorset to Yorkshire in England, and as scattered outliers in central and southern Wales, upon the Hebridean islands off northwest Scotland, in northeast Scotland and around the edges of the Northern Ireland province (Morton *in* Cope *et al.* 1980a; Fig. 1.1). They also occur at depth in boreholes in the southern Irish Sea (Celtic Sea), the English Channel, and in subcrop below the late Mesozoic and Tertiary of southeast England. Jurassic strata are present in boreholes in the Sole Pit region of the southern North Sea, in the northern North Sea and Moray Firth region off the east coast

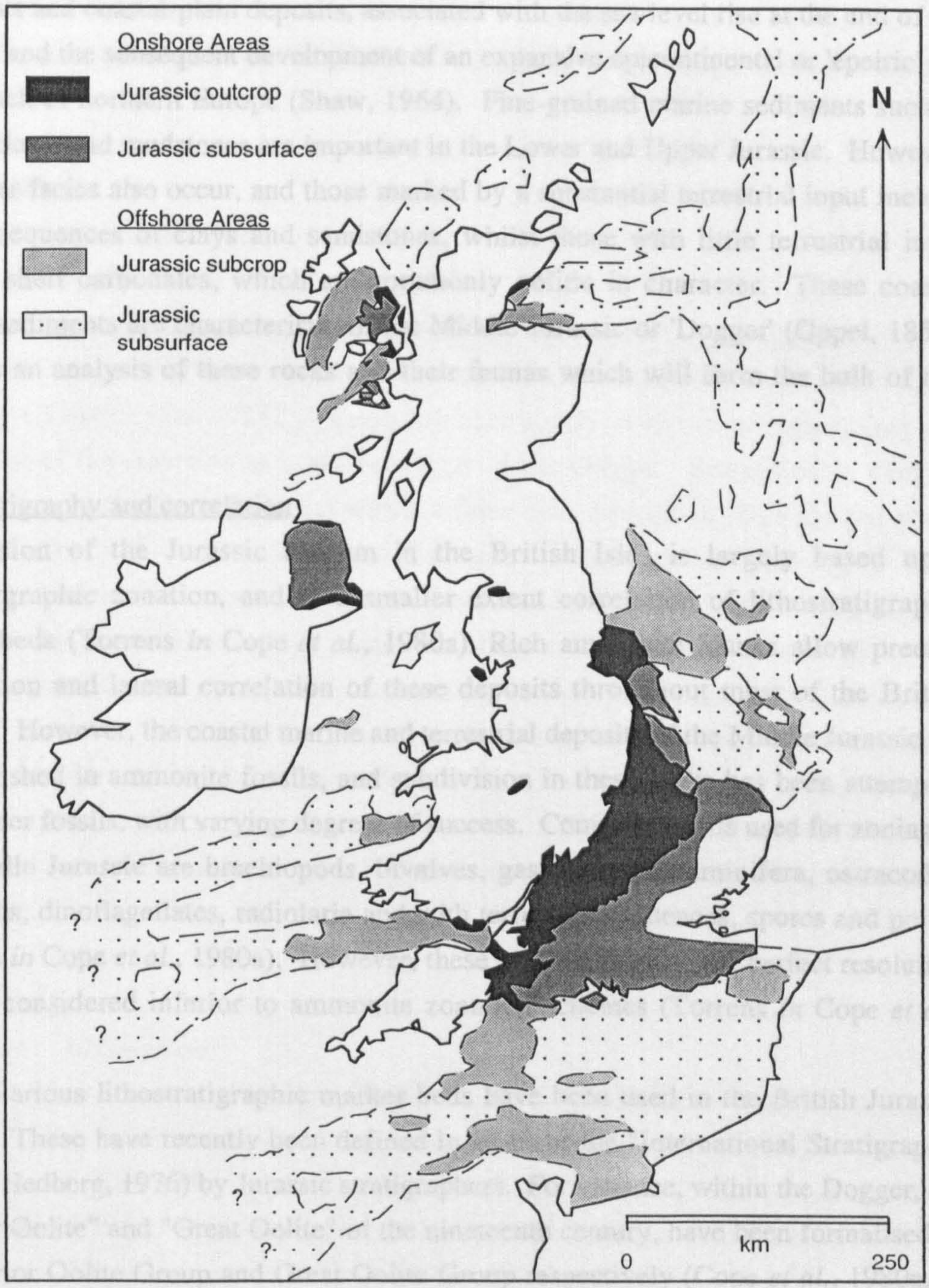


Figure 1.1. The sedimentary outcrop and possible extent of subcrop of the Jurassic System in the British Isles. North (N) is indicated.

of Scotland, and in the Minch basin off the Scottish west coast (Fig. 1.1). Offshore and onshore Jurassic rocks are important oil and gas reservoirs, and the succession contains many good source rocks.

The Jurassic System in the British Isles is composed of shallow marine sediments and coastal-plain deposits, associated with the sea-level rise at the end of the Triassic and the subsequent development of an expansive epicontinental or 'epeiric' sea over much of northern Europe (Shaw, 1964). Fine-grained marine sediments such as shales, clays, and mudstones are important in the Lower and Upper Jurassic. However, shallower facies also occur, and those marked by a substantial terrestrial input include deltaic sequences of clays and sandstones, whilst those with little terrestrial input include shelf carbonates, which are commonly oolitic in character. These coastal marine sediments are characteristic of the Middle Jurassic or 'Dogger' (Oppel, 1856), and it is an analysis of these rocks and their faunas which will form the bulk of this thesis

1.2. Stratigraphy and correlation

Subdivision of the Jurassic System in the British Isles is largely based upon biostratigraphic zonation, and to a smaller extent correlation of lithostratigraphic marker beds (Torrens *in* Cope *et al.*, 1980a). Rich ammonite faunas allow precise subdivision and lateral correlation of these deposits throughout most of the British Jurassic. However, the coastal marine and terrestrial deposits of the Middle Jurassic are impoverished in ammonite fossils, and subdivision in these rocks has been attempted using other fossils, with varying degrees of success. Common fossils used for zoning in the Middle Jurassic are brachiopods, bivalves, gastropods, foraminifera, ostracodes, coccoliths, dinoflagellates, radiolaria and with terrestrial sequences, spores and pollen (Torrens *in* Cope *et al.*, 1980a). However, these give much less than perfect resolution and are considered inferior to ammonite zonation schemes (Torrens *in* Cope *et al.*, 1980a).

Various lithostratigraphic marker beds have been used in the British Jurassic System. These have recently been defined in terms of the "International Stratigraphic Guide" (Hedberg, 1976) by Jurassic stratigraphers. For instance, within the Dogger, the "Inferior Oolite" and "Great Oolite" of the nineteenth century, have been formalised as the Inferior Oolite Group and Great Oolite Group respectively (Cope *et al.*, 1980a, b, and references therein). This thesis follows the lithostratigraphic nomenclature for the Dogger proposed by Cope *et al.*, (1980b), with modifications made since that work referenced by the author in the text. The biostratigraphic subdivision of the Dogger employed here is based upon that used in Cope *et al.* (1980b) and Harland *et al.* (1991), with changes referred to in the text.

The British Dogger is split into four stages : the Aalenian, the Bajocian, the Bathonian, and the Callovian (Cope *et al.*, 1980b, Harland *et al.*, 1991 and references therein; Fig. 1.2). Arkell (1946) was first to subdivide the British Middle Jurassic, based upon von Buch's (1839) division of the German Jurassic. He chose the base of the Middle Jurassic as the base of the Bajocian (*sensu anglico,* i.e. encompassing the Aalenian), and the base of the Oxfordian as the base of the Upper Jurassic (Arkell, 1946).

1.2.1. Aalenian

Basal Zone : Opalinum Zone

Basal Subzone : Opalinum Subzone

The name Aalenian was proposed for the first stage of the Dogger by Mayer-Eymar in 1864 working in the northern Alps, around the village of Aalen-Attenhofer, in Germany (Harland *et al.*, 1991). There has been much controversy surrounding the acceptance of the Aalenian as a separate stage of the Dogger. Some British workers consider it to be the Early Bajocian within a three-fold division of the Bajocian stage, namely Morton (1971) and others at the Colloque Jurassique in Luxembourg in 1967, although, at an earlier meeting (the Colloque Jurassique of 1962) the voting figures showed a majority in favour of the separation of the Aalenian stage (Mauberge, 1964). The latter practice has gained wide acceptance in Europe, where the Aalenian and Bajocian strata are lithologically quite distinct, and it is now putatively accepted in Britain, where the lithological distinction is not so clear-cut, although faunal division between the two stages is accepted (e.g. Cope *et al.*, 1980b; Fig. 1.2a). Subdivision into a scheme of standard ammonite zones is based upon work carried out by Contini *et al.* (1971) (Fig. 1.2a).

1.2.2. Bajocian

Basal Zone : Discites Zone

The Bajocian stage was proposed for strata outcropping around the town of Bayeux, France, by d'Orbigny in 1852, where the base of the stage is taken as the base of the *discites* Zone (Morton, 1971, Harland *et al.*, 1991; Fig. 1.2). The standard zonation of the Bajocian shown here in Figure 1.2b., is taken from Parsons (1976a, 1976b).

1.2.3. Bathonian

Basal Zone : Zigzag Zone

Basal subzone : Convergens Subzone

The Bathonian stage is named for the rocks which outcrop around the city of Bath, southwest England. However, the type section for the initial boundary is at Bas

a

	ZONES	SUBZONES
AALENIAN	*Graphoceras concavum*	*Graphoceras formosum horizon*
		G. concavum
	Ludwigia murchisonae	*Brasilia gigantea horizon*
		Brasilia bradfordensis
		L. murchisonae
		L. haugi
	Leioceras opalinum	*Tmetoceras scissum*
		L. opalinum

b

		ZONES	SUBZONES
BAJOCIAN	UPPER	*Parkinsonia parkinsoni*	*P. bomfordi*
			Strigoceras truellii
		Strenoceras (Garantiana) garantiana	*P. acris*
			St. (Garantiana) tetragona
			St. (G.) subgaranti
			St. (Pseudogarantiana) dichtoma
		Strenoceras subfurcatum	*St. (G.) baculata*
			Caumontisphinctes polygyralis
			Teloceras banksi
	LOWER	*Stephanoceras humphriesianum*	*T. blagdeni*
			S. humphriesianum
			Dorsetensia romani
		Emileia (Otoites) sauzei	
		Witchelia laeviuscula	*W. laeviuscula*
			Sonninia (Fissilobiceras) ovalis
		Hyperlioceras discites	

c

		ZONES	SUBZONES
BATHONIAN	UPPER	*Clydoniceras (Clydoniceras) discus*	*C. (C.) discus*
			C. (C.) hollandi
		Oppelia (Oxycerites) aspidoides	
		Procerites hodsoni	
	MIDDLE	*Morrisiceras (Morrisiceras) morrisi*	
		Tulites (Tulites) subcontractus	
		Procerites progracilis	
	LOWER	*Asphinctites tenuiplicatus*	
		Zigzagiceras (Zigzagiceras) zigzag	*Oppelia (Oxycentes) yeovilensis*
			M. (Morphoceras) macrescens
			P. (Parkisonia) convergens

d

	ZONES	SUBZONES
CALLOVIAN	*Quenstedtoceras (Lamberticeras) lamberti*	*Q. (L.) lamberti*
		Q. (Eboraciceras) henrici
	Peltoceras athleta	*Kosmoceras (Kosmoceras) spinosum*
		K. (Labokosmokeras) proniae
		K. (L.) phaeinum
	Erymnoceras coronatum	*K. (Zugokosmokeras) grossouvrei*
		K. (Z.) obductum
	Kosmoceras (Gulielmites) jason	*K. (G.) jason*
		K. (G.) medea
	Sigaloceras calloviense	*S. (Catasigaloceras) enodatum*
		S. calloviense
		Proplanulites koenigi
	Macrocephalites (Macrocephalites) macrocephalus	*M. (Kamtokephalites) kamptus*
		M. (M.) macrocephalus

Figure 1.2. Biostratigraphic columns for the four stages of the Dogger : (a) the Aalenian (the oldest); (b) the Bajocian; (c) the Bathonian and (d) the Callovian (the youngest).
Biostratigraphic subdivision is by ammonite zones and subzones, after Cope *et al.* 1980b.

Auran, in the Basses-Alpes, in southeast France (Harland *et al.*, 1991). Torrens noted (*in* Cope *et al*, 1980b, p.21) that the extreme rarity and in some cases, complete absence of ammonites within many British Bathonian deposits, has presented many problems in establishing a scheme of standard zonation in the stage, and perhaps more importantly, in correlating sections within this standard. This thesis follows the zonation scheme based on those in Torrens (1974), Cope *et al.* (1980b)(Fig. 1.2c).

1.2.4. Callovian

Basal Zone : Macrocephalus Zone

Basal Subzone : Mcrocephalus Subzone

The Callovian stage was named from the occurrence of the Kellaways Beds within the lower part of the southern English succession (Harland *et al.*, 1991). The basal boundary was proposed in the Chippenham area of Wiltshire, as the base of the Upper Cornbrash (Morton, 1971, Duff *in* Cope *et al.*, 1980b; Fig. 1.2d). Subdivision of the Callovian by standard zonation, is based upon that in Callomon (1964), with modifications by Duff (*in* Cope *et al.*, 1980b) (Fig. 1.2d).

1.2.5. Chronostratigraphy of the Jurassic (Table 1.1)

The geologic time scale for the Jurassic Period is illustrated here (Table 1.1, Fig. 1.3). Ages of the stage boundaries are taken from the most recent reviews of the Jurassic geological time scale and are recorded in millions of years before present. Table 1.1 lists the four most recent chronostratigraphic time scales based upon an integration of radiometric dating techniques, biostratigraphy and magnetostratigraphy, which are those by van Hinte (1978), Harland *et al.* (1982, 1991) and Haq *et al.*, (1987, 1988). These timescales are reviewed below and their merits considered. However, any further references to dates and ages of system boundaries in the text, will refer to the "Standard Global Geochronostratigraphic Time Scale, 1989" given by Harland *et al.* (1991), for example Fig. 1.3.

Radiometric dating analyses of whole rock and minerals such as biotite and feldspar, provide the first standard by which stage boundaries are dated. The radiogenic isotopic dating techniques most used in the Jurassic, are based on Potassium-Argon (K-Ar) and Rubidium-Strontium (Rb-Sr) radioactive isotopic decay rates (Harland *et al.*, 1991). These techniques are sometimes called 'absolute dating', although Harland *et al.* (1991) prefer the term 'chronometric' techniques. The geological chronometric scale is the standard linear time scale, made up of equal time units (seconds or years) counted before the present (B.P.; before 1950, Harland *et al.*, 1991; Table 1.1). Problems with radiometric dating techniques and the accuracy of the chronometric time scale are reviewed by Haq *et al.* (1987) and Harland *et al.* (1982, 1991). There are about nine

	CHRONOSTRATIGRAPHIC TIME SCALES /Ma				CHRONOMETRIC AGE DATA (Harland 89)		
STAGE BOUNDARY	van Hinte 78	Harland 82	Haq 87, 88	Harland 89	C-gram 89	Interpolated C-gram 89	"Tie-points" 89
Berriasian-Tithonian	135.0	144	131	145.6	148.0	145.0	145
Tithonian-Kimmeridgian	141.5	150	136	152.1	151.0	153.5	156
Kimmeridgian-Oxfordian	145.0	156	145	154.7	156.0	156.0	
Oxfordian-Callovian	151.5	163	152	157.1	156.3	159.8	
Callovian-Bathonian	158.5	169	157	161.3	159.0	162.7	
Bathonian-Bajocian	165.0	175	165	166.1	159.2	166.6	
Bajocian-Aalenian	170.0	181	171	173.5	177.0	173.5	(168)
Aalenian-Toarcian	175.0	188	179	178.0	178.0	178.0	
Toarcian-Pliensbachian	180.0	194	186	187.0	182.0	187.0	
Pliensbachian-Sinemurian	185.0	200	194	194.5	189.5	194.5	
Sinemurian-Hettangian	190.0	206	201	203.5	203.5	203.5	203.5
Hettangian-Rhaetian	195.0	213	210	208.0	210.5	208.0	209.5

Table 1.1. Chronostratigraphic time scales and chronometric age data for the Jurassic, after van Hinte (1978), Harland (1982, 1989), and Haq *et al.* (1987, 1988).

well-defined radiometric dates within the Dogger (van Hinte, 1978, Harland *et al.*, 1991).

Producing a Jurassic chronometric time scale is made more difficult by the fact that there is no accurate magnetostratigraphy for the period. The magnetostratigraphic timescale is based upon ascertaining geomagnetic polarity reversals in the Earth's magnetic field, and the assumption of a constant sea-floor spreading rate throughout geological time. It is measured within oceanic crustal and sea-floor rocks, and has provided an independent means of measuring 'absolute' dates (van Hinte, 1978). However, for rocks older than the Callovian-Oxfordian it is of limited use because sea-floor magnetic anomaly data for rocks pre-Oxfordian age is extremely rare. The Triassic-Callovian magnetostratigraphic schemes given by Haq *et al.* (1987, 1988) and in Harland *et al.* (1982, 1991) are based upon various palaeomagnetic studies of outcrops and sections on land, from which the data have been modelled into a synthetic scheme for the Earth's magnetic polarity reversals.

The main problem with all the chronostratigraphic time scales shown in Table 1.1 is the difficulties encountered in trying to provide a straight correlation between the biostratigraphic column with ages given by radiometric dating and the Jurassic magnetostratigraphic schemes. The correlation of geomagnetic polarity with the geochronostratigraphic time scale for the Jurassic has not been attempted by many authors (e.g. Harland *et al.*, 1991). However, notable exceptions include the use of geomagnetic anomalies to tie in chronostratigraphy with cycles of sea-level change in the Mesozoic, by Vail & Todd (1981), Vail *et al.* (1984), and Haq *et al.* (1987, 1988).

Until correlation of these factors has been better constrained, the Jurassic chronostratigraphic time scales (Table 1.1) have had to rely upon assigning standard equal durations to the subdivisions of the biostratigraphic zonal scheme (based upon ammonites), within the individual stages of the Jurassic System (Haq *et al.*, 1987). The linear chronostratigraphic time scales given by van Hinte (1978) and Haq *et al.* (1987, 1988) are based upon assigning an equal duration of approximately one million years to an ammonite zone (Table 1.1; the duration of Jurassic ammonite zones is given as between 0.96-1.15 Ma by van Hinte (1978), based upon work by Arkell, 1956). Hence, the length of a stage is equal to the number of zones it contains (van Hinte, 1978). These were then set between several well-constrained radiometric or magnetic dates at stage boundaries, to produce a linear time scale. The time scale of Harland *et al.* (1982, 1991) is similar in relying upon a series of well dated 'tie-point' ages, which are also placed at chronostratic boundaries such as zone or stage boundaries (Table 1.1, Harland *et al.*, 1991).

From the radiometric data tie-point ages, Harland *et al.* (1982, 1991) have produced a suite of computer-aided graphs, known as 'chronograms' or 'C-grams', to work out the best-fit date for each stage boundary and the standard error involved

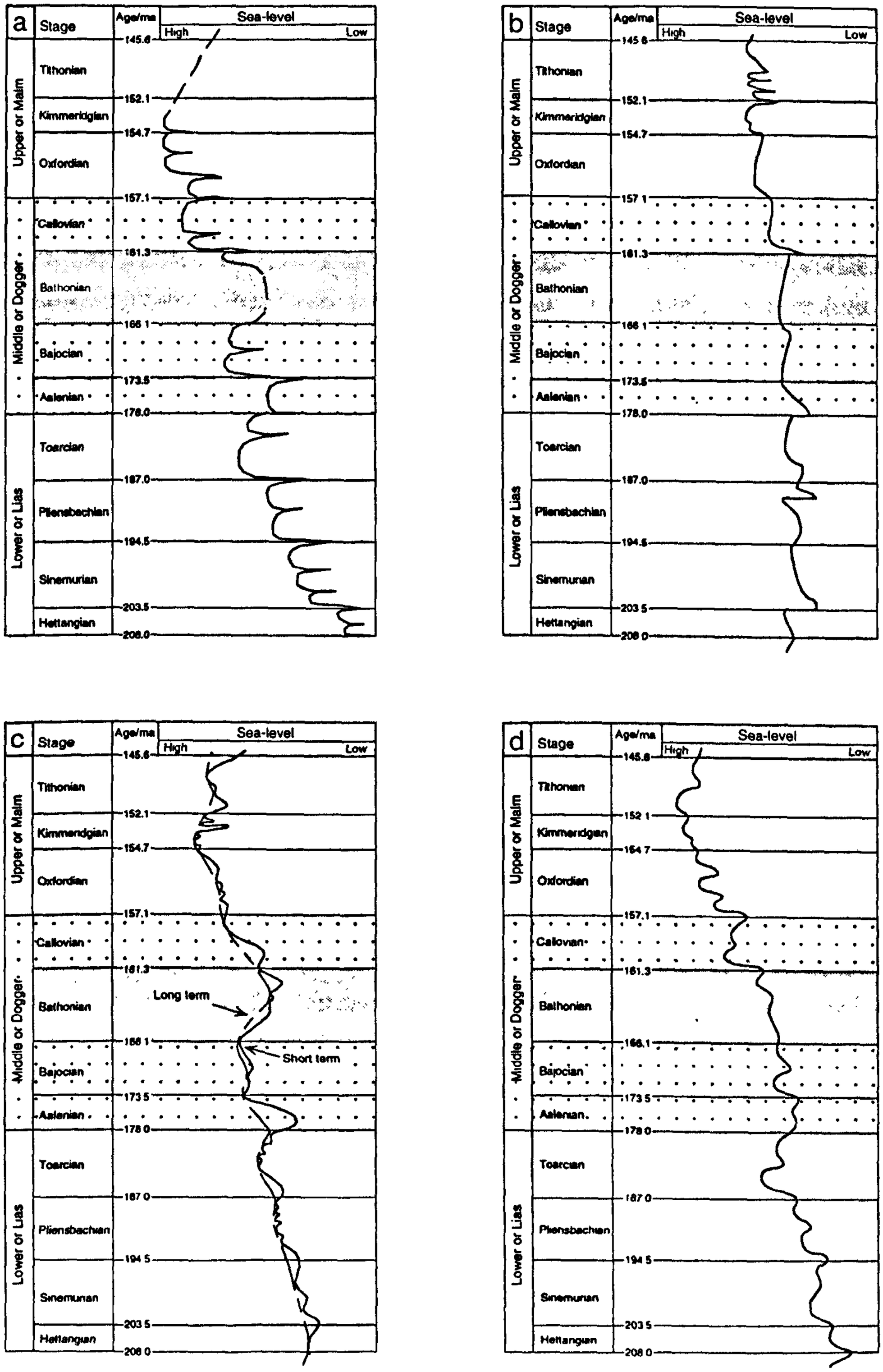


Figure 1.3. Sea-level curves for the Jurassic, based upon (a) Hallam, 1978, (b) Vail *et al.*, 1977, Vail & Todd, 1981, (c) Vail *et al.*, 1984, Haq *et al*, 1987, 1988, and (d) Hallam, 1988.

(Table 1.1., after Harland *et al.*, 1991). Interpolated ages of these boundaries are based upon the results obtained from the chronogram analysis and are set within the biostratigraphic scheme of ammonite zones (or 'chrons', Harland *et al.*, 1982, based on the scheme in Cope *et al.*, 1980a,b; Table 1.1.), were geological time is proportional to the number of zones between the tie-points. This statistical method by which the suitability of any age value for a particular stratigraphic boundary is evaluated, is described in Harland *et al.* (1991, chapter 5). However the "Standard Global Geochronostratigraphic Time Scale, 1989" defined by Harland *et al.* (1982, 1991) is not an 'equal stage' scheme. For example, within the "Standard Global Geochronostratigraphic Time Scale, 1989", the Sinemurian stage of the Lower Jurassic has 6 ammonite zones, but is 9 million years in duration, whereas the Bathonian stage is 4.8 million years long and yet contains 8 ammonite zones (Harland *et al.*, 1991; Table 1.1., Fig. 1.2).

1.2.6. The Jurassic sea-level curve (Fig. 1.3.)

The state of sea-level or 'eustasy' in the Jurassic has been worked out by studying the facies changes in coastal sequences and biostratigraphic data at outcrop to evaluate the change through time of the areal extent of world-wide epicontinental seas (for example, Hallam, 1978, 1981, 1988) or by synthesising data taken from seismic sections to investigate the degrees of global coastal onlap (for example, Vail *et al.*, 1977, 1984, Vail & Todd, 1981, Haq *et al.*, 1987, 1988). Figure 1.3. illustrates the four eustatic curves :

(1) The first sea-level curve for the Jurassic (Fig. 1.3a; after Hallam, 1978) is based upon the integration of global transgressive and regressive events as seen in the rock record at outcrop and by basin analysis, with a stage-by-stage plot of the areal spread of Jurassic marine deposits (as explained in Hallam, 1981). It shows a general sea-level rise over the Jurassic period - a trend known as "positive eustasy" (Hallam, 1978, p.1), which is seen in all the eustatic curves for the Jurassic (Fig. 1.3). The general secular trend for the Jurassic proposed by Hallam (1978, 1981) is a long term rapid increase in sea-level in the early Jurassic, followed by a long period of stillstand in the Middle Jurassic up until the Oxfordian-Kimmeridgian, followed by a rapid drop in eustasy in the late Jurassic (Tithonian). Hallam (1978) estimates that the overall rise in sea-level from the Rhaetian to the late Jurassic was about 150 metres.

The curve shows a 'pulsed' behaviour of short-term transgressive and regressive phases. Regressions appear almost instantaneous in character, which may be an artefact of the rock record, as facies changes within sedimentary sequences show that deepening events appear much sharper than those indicating shallowing events, and regressive sequences may also be condensed. Therefore, the most precise determination of the age and length of a given shallowing event can only be obtained from areas with the most

complete stratigraphic successions. This pulsed behaviour was remarked and improved upon in later Hallam sea-level curves (for instance, Hallam, 1981) and his updated eustatic curve for the Jurassic shows a much more sinusoidal shape. The short-term pulses upon Hallam's curve indicate important sea-level rises in the early Hettangian, the early-mid Toarcian, the early Bajocian, the late Bathonian-early Callovian and the mid Oxfordian. Important sea-level falls are in the late Toarcian-early Aalenian, the early Bathonian, the late Callovian and the late Tithonian.

(2) The first curve produced by seismic stratigraphy for the Jurassic system, was attempted by Vail *et al* (1977) and reproduced with some modifications in Vail & Todd (1981), this is shown in Figure 1.3b. The eustatic curve was produced by assessing seismic sections and well-log data for offshore Jurassic basins, and was largely the work of the Exxon oil company research team - hence the term "Exxon curve" for this type of sea-level plot (Hallam, 1988). Global sea-level fluctuations are based upon analysis of large and medium scale unconformities (which appear as major reflectors in seismic sections) to assess the degrees of coastal onlap within a given basin (Vail & Todd, 1981). Hallam (1981, 1988) has criticised this technique for relying upon classified and therefore untestable oil company data, and for being based largely upon northern European, and in particular North Sea, seismic stratigraphy.

The general eustatic pattern shown in Figure 1.3b is almost identical to the plot for coastal onlap shown for the North Sea Jurassic (Vail & Todd, 1981; and remarked upon in Hallam, 1981, 1988). The characteristic "dog-tooth" pattern of rapid, almost instantaneous sea-level fall is almost certainly a function of this. However, following major revisions, the final sea-level curve produced by the Exxon team shows a much smoother profile (Fig. 1.3c; after Vail *et al.*, 1984, Haq *et al.*, 1987, 1988). The general long-term trend of sea-level fluctuations is again one of secular sea-level rise throughout the Jurassic Period, with an overall increase of around 75-100 metres (Vail & Todd, 1981, Vail *et al.*, 1984). The long term trend as described in Figure 1.3c is a slow increase in sea-level through the early Jurassic, with a marked dip in the late Hettangian. Sea-levels remained low throughout much of the Middle Jurassic, and are particularly low in the Aalenian and Bathonian, although there is a major transgressive phase in the Bajocian. In the Callovian there was a long-term eustatic rise in sea-level which progressed throughout the rest of the period, rising to a peak in the Kimmeridgian, before a rapid reversal and decline in sea-level in the Tithonian-Berriasian (Haq *et al.*, 1987). This long term sea-level curve for the Jurassic is known as the "Jurassic supercycle" (Vail *et al.*, 1977).

Superimposed upon the long-term sea-level curve are around 13 shorter-term eustatic cycles (Vail *et al.*, 1984). These are seen as major unconformities upon seismic sections, and are interpreted as global stratigraphic events (Vail *et al.*, 1984). Of these, seven are subaerial and submarine unconformities known as "Type-One

Unconformities" (Vail *et al.*, 1977), which are believed to be caused by rapid eustatic falls in sea-level. These include regressions in the latest Hettangian, the late Pliensbachian, the early Aalenian, the late Kimmeridgian, and three events within the Tithonian sequence (Fig. 1.3c; Vail *et al.*, 1984). Six unconformities are subaerial only and are known as "Type-Two Unconformities" (Vail *et al.*, 1977). These occur within the basal Pliensbachian, the early Toarcian, the late Bajocian, the late Bathonian, the early Callovian and the late Oxfordian and are believed to be related to slow eustatic falls in sea-level (Fig. 1.3c; Vail *et al.*, 1984). In addition the Jurassic succession shows 14 condensed sequences which indicate rapid eustatic sea-level rises, these occur in the late Sinemurian, the late Pliensbachian, the basal Toarcian, the mid Toarcian, the basal Bajocian, the basal Bathonian, the basal Callovian, the mid Callovian, the mid Oxfordian, the basal Kimmeridgian, the basal Tithonian, two more within the Tithonian and a basal Berriasian event (Fig. 1.3c; Vail *et al.*, 1984).

(3) Following the publication of the revised "Exxon curve" (Vail *et al.* 1984, Haq *et al.*, 1987), Hallam generated a new eustatic sea-level curve for the Jurassic based upon a revaluation of his earlier curves and those of the Exxon team (Hallam, 1988). Hallam (1988) estimates overall secular rise in sea-level for the Jurassic to be about 100 metres. The resulting curve (Fig. 1.3d) is quite similar in general trend and sinusoidal pattern of sea-level fluctuations, to that of the Exxon team (Fig. 1.3c), and was produced to test the validity of both sets of analytical data. Both curves show important transgressive events in the Sinemurian, Toarcian, Bajocian, Callovian, Oxfordian and Kimmeridgian and Hallam (1988) notices 18 shallowing/deepening events compared to the 17 noticed by Vail *et al.* (1984) within the Jurassic supercycle.

However, there are some major deviations between the two final Jurassic eustatic curves (Figs. 1.3c and 1.3d). For instance, Hallam (1981, 1988) recognises a major unconformity at the base of the Hettangian, and he states that global stratigraphic evidence indicates that the lowstand in the Lower Jurassic occurred at the beginning of the Hettangian (Fig.1.3d; Hallam, 1988) and not within the early Sinemurian as indicated by the Exxon curve (Fig. 1.3c). Other differences are less pronounced, for instance the larger and more rapid transgressive events indicated in Hallam's curve (Fig. 1.3d) for the basal Sinemurian, early Toarcian and early Callovian. However, most importantly Hallam (1988) considers the large regression indicated within the Aalenian and the "noisy" Kimmeridgian-Tithonian part of the Exxon curve (Fig. 1.3c) are the result of localised tectonic and sedimentation fluctuations in the North Sea basins on which the Exxon curve is largely based and therefore not globally significant (Hallam, 1988, p.271).

In summary, the general global eustatic trend for the Jurassic supercycle is one of increasing sea-level, following the very low sea-levels in the late Palaeozoic (Haq *et al.*, 1987). This trend was checked and reversed some time in the Tithonian.

Many theories have been proposed for the controls for eustatic sea-level fluctuations (see review in Hallam, 1988). The two most likely mechanisms are changes in the amount of polar ice and changes in the volume of the ocean basins (Anderton *et al.*, 1979). As there is no evidence for Jurassic polar ice-caps, the most widely favoured mechanism is the changing volume of oceanic ridge systems producing volumetric changes within the oceans (Hallam, 1978). Hence, transgressive phases are produced during periods of accelerated sea-floor spreading or by the production of new tracts of the ridge system. The global rise in eustasy in the middle Mesozoic may be related to the break-up of the supercontinent Pangaea and the Jurassic supercycle could be related to the onset of central Atlantic sea-floor spreading in the late Triassic-early Jurassic (Pitman & Talwani, 1972).

Hallam (1978, 1988) points out that plate boundary motions are not continuous, but occur in a series of pulses. He suggests that the smaller-scale fluctuations in sea-level occur in a series of cycles with an average duration of around four million years (Figs. 1.3c and 1.3d), and estimates a rate of eustatic change comparable to the small scale growth and decay of oceanic ridge systems (i.e. around a few centimetres per thousand years; Hallam, 1988, after Pitman, 1978).

1.3. The Middle Jurassic World

Following the onset of north Atlantic rifting, hailing the fragmentation of the Upper Palaeozoic supercontinent "Pangaea" in the late Triassic and early Jurassic (Pitman & Talwani, 1973), marine conditions spread over much of the low-lying marginal alluvial plains during the Rhaetian transgression (latest Triassic). By the early Jurassic, marine transgression had transformed much of northern Europe into an enormous, shallow epeiric sea. The extensive shelf sea was protected from strong tidal and storm influences, and distinctive facies of laminated bituminous shales, and rhythmic sequences of lime mud and marl accumulated during much of the Jurassic. Over more shallow regions (such as intrabasinal swells), oolitic ironstones and condensed cephalopod limestones, such as the famous Ammonitico Rosso of southern Europe, developed in the relative absence of any terrigenous input. The shelf sea gave way in the south to the open oceanic deposits of the Tethys Ocean (Fig. 1.4).

The marine conditions in northern Europe continued throughout Jurassic time with two major regressive intervals, one during the Middle Jurassic and one in the latest Jurassic (Tithonian-Berriasian; Fig. 1.3). These regressive periods can be attributed to a combination of eustatic sea-level fall and local tectonism due to the Atlantic rifting events (Pitman & Talwani, 1973, Anderton *et al.*, 1979). Toward the end of Liassic

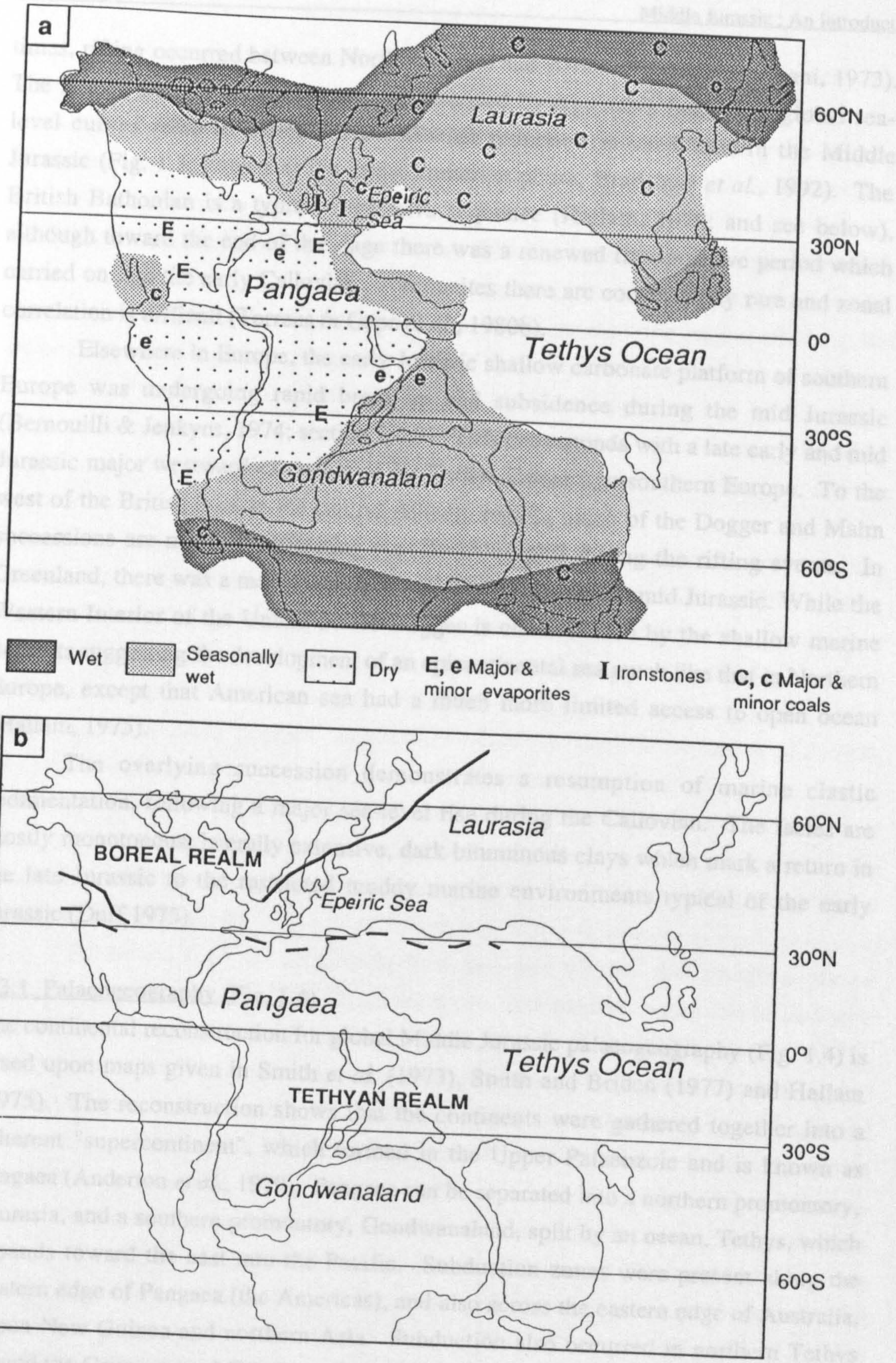


Figure 1.4. The palaeogeography of the world in Middle Jurassic time, showing (a) the precipitation belts upon the Pangaean continental mass and sites of major coal, evaporite, and ironstone deposits, after Hallam, 1984, 1985; and (b) the location and extent of the oceanic faunal realms : Boreal and Tethyan, after Hallam, 1975.

times, rifting occurred between North America and Africa (Pitman & Talwani, 1973). The British Dogger sequences are characterised by regressive facies and global sea-level curves seem to indicate a world-wide reduction in shelf seas in the Middle Jurassic (Fig. 1.3; known as the mid-Cimmerian phase, Bradshaw *et al.*, 1992). The British Bathonian is a typical regressive sequence (Hallam, 1978; and see below), although toward the end of the stage there was a renewed transgressive period which carried on into the early Callovian. Ammonites there are consequently rare and zonal correlation is difficult (Torrens *in* Cope *et al.*, 1980b).

Elsewhere in Europe, the early Jurassic shallow carbonate platform of southern Europe was undergoing rapid break-up and subsidence during the mid Jurassic (Bernouilli & Jenkyns, 1974; section 1.3.3). This corresponds with a late early and mid Jurassic major westward extension of the Tethys Ocean into southern Europe. To the west of the British Isles in the central Atlantic region, much of the Dogger and Malm successions are missing suggesting considerable uplift during the rifting events. In Greenland, there was a major marine transgression during the mid Jurassic. While the Western Interior of the United States Dogger is characterised by the shallow marine deposits suggesting the development of an epicontinental sea much like that in Northern Europe, except that American sea had a much more limited access to open ocean (Hallam, 1975).

The overlying succession demonstrates a resumption of marine clastic sedimentation, following a major sea-level rise during the Callovian. The facies are mostly monotonous, laterally extensive, dark bituminous clays which mark a return in the late Jurassic to the restricted muddy marine environments typical of the early Jurassic (Duff 1975).

1.3.1. Palaeogeography (Fig. 1.4)

The continental reconstruction for global Middle Jurassic palaeogeography (Fig. 1.4) is based upon maps given in Smith *et al.* (1973), Smith and Briden (1977) and Hallam (1975). The reconstruction shows that the continents were gathered together into a coherent "supercontinent", which formed in the Upper Palaeozoic and is known as Pangaea (Anderton *et al.*, 1979). Pangaea can be separated into a northern promontory, Laurasia, and a southern promontory, Gondwanaland, split by an ocean, Tethys, which expands toward the east into the Pacific. Subduction zones were present along the western edge of Pangaea (the Americas), and also across the eastern edge of Australia, Papua New Guinea and northern Asia. Subduction also occurred in northern Tethys around the Caucasus and Crimea regions (Hallam, 1975).

During the Middle Jurassic, the British Isles lay at about 35°N, some 10°S of their present position. Britain occupied an area forming part of the northern European epeiric sea, in a region of overlap between the Tethyan and Boreal faunal realms (Fig.

1.4b). The Fenno-Scandinavian shield was the nearest landmass of any size (Sellwood & Jenkyns, 1975). The palaeogeographic maps for the four stages of the Dogger are shown in Fig. 1.5, taken from the most recent review of British Jurassic geography by Bradshaw *et al.* (1992). As Dogger rocks cover only about 25% of the map areas and are absent from many of the western basins and structural highs (Bradshaw *et al.*, 1992), hence the positions and interpretation of land and sea areas are fairly conjectural.

Landscape relief appears to have been fairly low, with the highest landmasses, the Scottish, Scandinavian and Central North Sea High, attaining heights of about 500 metres above sea-level (Bradshaw *et al.*, 1992). More subdued features such as the London-Brabant massif, Cornubian and Welsh landmass may have only been a couple of hundred metres in height at most. The landmasses and islands were fringed by coalescent deltas, coastal marshes and low-lying alluvial plains building out into the shallow sea.

The London-Brabant massif and Cornubian island formed the major landmasses in the south of England, and these pass to the south into the marine conditions of the Weald, Channel and Paris basins (Megnien & Megnien, 1980; Fig. 1.5). To the west and south of Cornubia, the Celtic Sea Basin and Western Approaches Trough are also fully marine basins, which open to the south into the Bay of Biscay Trough. The Welsh landmass was a strong positive feature in mid Jurassic times (Bradshaw *et al.*, 1992), but whether it was connected to the Pennine High and whether the Pennines formed such a positive structure as they do at the present is a point of contention (Bradshaw *et al.*, 1992). However, between the Welsh, Pennine and London-Brabant landmasses the fairly shallow water marginal marine conditions of the Midland or Cotswold-Weald Shelf prevailed (Fig. 1.5)

To the north of the London-Brabant land area alluvial and marginal marine deposits were deposited into the Cleveland Basin. Sedimentation into this basin seems to have been from the Scottish landmass and Central North Sea High (Thomas, 1975). Volcanic doming in the junction between three structural grabens (the Viking and Central Grabens to the north, and the Moray Firth Basin to the south) also contributed to the high sedimentation rates in the northern North Sea. The volcanic centre was initiated in the late Toarcian-early Aalenian (Howitt *et al.*, 1975; Fig. 1.5a) and appears to be of subaerial fissure-type, extruding volcaniclastic material over a lava plateau of around 12 000km^2. Up to 4000 metres of volcanic deposits (Rattray Formation) are recorded in the area, which must have provided the surrounding non-marine basins with substantial sedimentation (Woodhall & Knox, 1979). The volcanicity waned and the dome began to subside in Mid to Late Bathonian times (Dixon *et al.*, 1981; Fig. 1.5d).

To the north of the Scottish landmass, it appears that the Shetland Platform was emergent (Richards *et al.*, 1988), but the presence of marine deposits within the Moray Firth Basin suggests that there was no connection between the two land areas (cf.

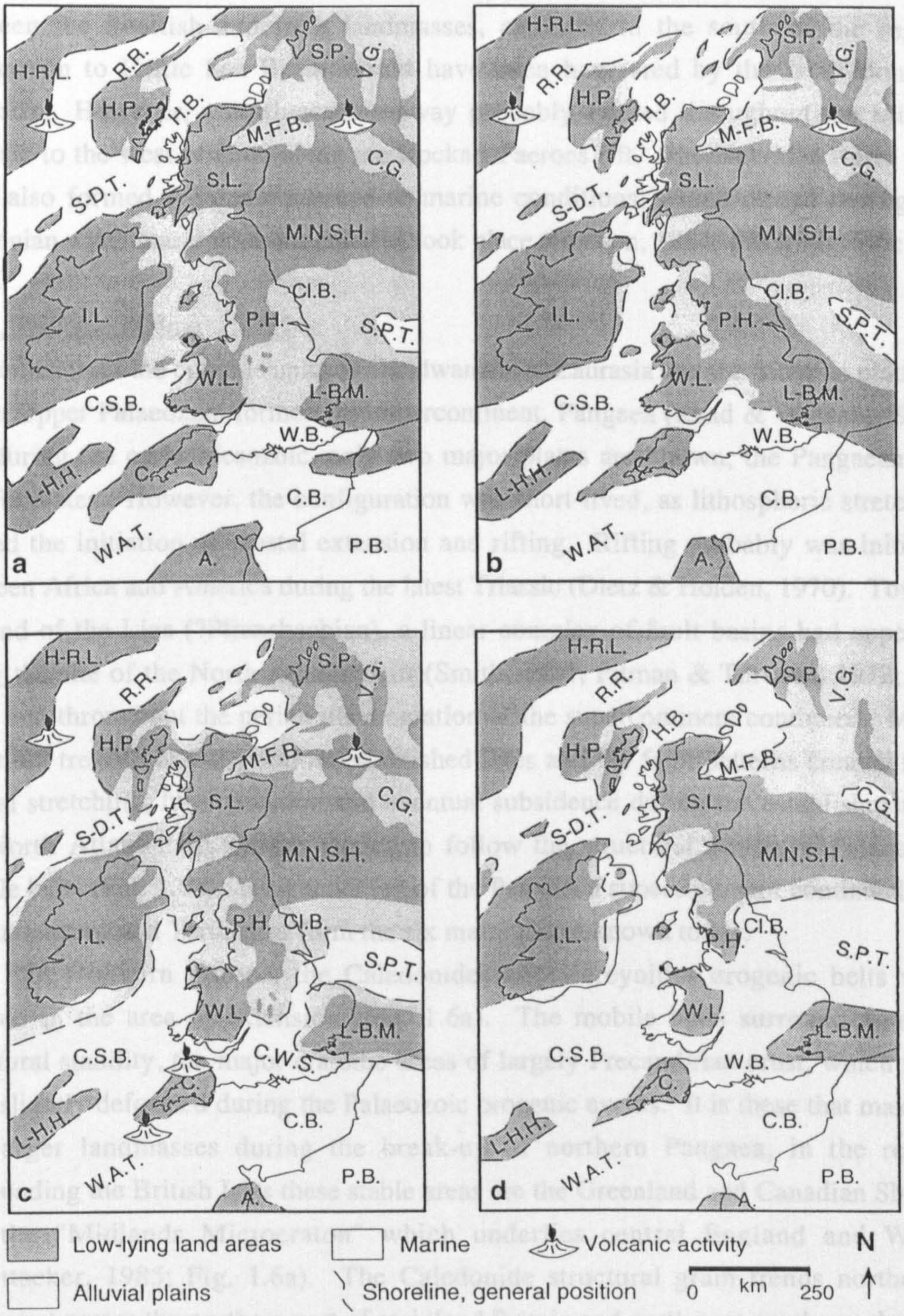


Figure 1.5. British palaeogeographic maps for the four stages of the Dogger : (a) Aalenian; (b) Bajocian; (c) Bathonian; and (d) Callovian, showing the position of land areas and shelf seas, after Bradshaw *et al.*, 1992. Landmass abbreviations: 'A.' Amorica, 'C.' Cornubia, 'H.P.' Hebrides Platform, 'H-R.L.' Hatton-Rockall Landmass, 'I.L.' Irish Landmass, 'L-B.M.' London-Brabant Massif, 'L-H.H.' Labadie Bank-Haig Fras Highs, 'M.N.S.H.' Mid North Sea High, 'P.H.' Pennine High, 'S.L.' Scottish Landmass, 'S.P.' Shetland Platform, 'W.L.' Welsh Landmass. Basinal area abbreviations: 'C.B.' Channel Basin, 'Cl.B' Cleveland Basin, 'C.S.B.' Celtic Sea Basins, 'C.G.' Central Graben, 'C.W.S' Cotswolds-Weald Shelf, 'H.B.' Hebrides Basin, 'M-F.B.' Moray-Firth Basin, 'P.B.' Paris Basin, 'R.R.' Rockall Rift, 'S-D.T' Slyne-Donegal Trough, 'S-P.T.' Sole Pit Trough, 'V.G.' Viking Graben, 'W.A.T' Western Approaches Trough, 'W.B.' Weald Basin. North (N) is indicated.

Sellwood & Jenkyns, 1975, Eynon, 1981). It is unclear whether there was a sea-way between the Scottish and Irish landmasses, certainly in the south of the region connection to Celtic Sea Basin would have been hampered by the Islay-Donegal Platform. However, a north-south seaway probably existed throughout the Middle Jurassic to the west of Eire, along the Rockall-Faeroes rift. The Hebrides Basin may have also formed a narrow passage to marine conditions, which closed during the Bathonian when marginal sedimentation took place (Hudson, 1964, 1983; Fig. 1.5c).

1.3.2. Tectonic regime

The collision of the cratonic units of Gondwanaland, Laurasia and the Siberian platform in the Upper Palaeozoic formed the supercontinent, Pangaea (Read & Watson, 1975), and during the early Mesozoic, only two major plates are known, the Pangaean and Pacific plates. However, the configuration was short-lived, as lithospheric stretching caused the initiation of crustal extension and rifting. Rifting probably was initiated between Africa and America during the latest Triassic (Dietz & Holden, 1970). Toward the end of the Lias (?Pliensbachian), a linear complex of fault basins had appeared along the site of the North Atlantic Rift (Smith, 1971, Pitman & Talwani, 1972; Fig. 1.6a) and throughout the period disintegration of the supercontinent continued. Major structural trends generally follow established lines and the fault patterns created from crustal stretching, fragmentation and eventual subsidence during the establishment of the North Atlantic rift system tended to follow the structural grains of Palaeozoic mobile belts (Fig. 1.6a). Fragmentation of the Pangaean supercontinent continued into the Cretaceous and Tertiary to form the six major plates known today.

In Northern Europe, the Caledonides and Hercynides orogenic belts were utilised in the area of extension (Fig. 1.6a). The mobile belts surround areas of structural stability, the major cratonic areas of largely Precambrian crust, which were only slightly deformed during the Palaeozoic orogenic events. It is these that make up the larger landmasses during the break-up of northern Pangaea, in the region surrounding the British Isles these stable areas are the Greenland and Canadian Shields and the "Midlands Microcraton" which underlies central England and Wales (Whittacker, 1985; Fig. 1.6a). The Caledonide structural grain trends northeast-southwest across the northern part of mainland Britain and northwest-southeast through the North Sea (Ziegler, 1981; Fig. 1.6a). The Hercynides follow an east-west trend in southern Eire, England and western Europe and intersect the older Caledonian belts (Chidlaw, 1987; Fig. 1.6a). On the Midlands Microcraton, a further north-south "Malvernoid" structural grain has also been noticed (Whittacker, 1985).

Tectonic systems involved in crustal tension are characterised by normal faulting, horst and graben block structures, long-term basinal subsidence and large-scale basic volcanics and intrusives (Hallam, 1975; Fig. 1.6). A horst-graben

Figure 1.6. The North Atlantic Rift system in the Lower and Middle Jurassic, showing pre-drift structural continuities of the Palaeozoic mobile belts and stable Precambrian cratonic areas across Laurasia and northern Gondwanaland (a), after Ziegler, 1981. North (N) is indicated. (b) Model to illustrate the North Sea basement horst and graben tectonics in the early and mid Jurassic and facies development across the basin and swell topography, after Hallam & Sellwood, 1976: 1) Deposition of silts and muds within an early Jurassic epeiric sea, over active basement faulting; 2) Middle Jurassic thermal doming in the North Sea, leads to uplift, regression and erosion at rift margins; 3) Domal collapse, tectonic quiescence and graben subsidence in the late Middle Jurassic, leads to a marine transgression and envelopment of much of the area in basinal shales and marginal sands around local positive regions.

a

Grabens

Basic volcanics

Basic dyke swarms

N

East Greenland Rift

CALEDONIDE

West Scottish Basins

North Sea Basins

LAURASIA

East Newfoundland Basin

Fundy Basin

Newark Group Basins

Midlands Microcraton

Rhine Graben

HERCYNIAN

APPALACHIAN

MAURATANIAN

Tethys Ocean

GONDWANALAND

1 Porcupine Trough
2 Celtic Sea Trough
3 Western Approaches Trough
4 Severn Basin
5 Rockall and north-west Irish Basins

Palaeozoic mobile belts (NAMED)

Grenvillian Belt (late Precambrian)

Lower Proterozoic

Archean

b

1

SL

2

Main sediment transport

Erosion

Erosion

Main sediment transport

SL

3

SL

Pre-Jurassic basement

Marine silts

Paralic sands

Bituminous shales, with basal trangressive sands

Marine/marginal sands and turbidites

SL Sea level

topography is suggested as an important factor in controlling sedimentation in the British Triassic (Audley-Charles, 1970a,b). Block-faulted systems are caused by thermally controlled crustal arching and fault collapse, due to lithospheric stretching in tensional regimes. These develop into a series of horsts and grabens, which form into tilted blocks with continued subsidence (Hallam & Sellwood, 1976; Fig. 1.6b).

A similar regime occurs in the Early Jurassic except that it is mostly marine sedimentation in the British Isles (Sellwood & Jenkyns, 1975) as many of the classic British Jurassic sedimentary basins correspond to Triassic graben structures (Fig. 1.6b). The Jurassic system in Great Britain is characterised by a number of condensed and expanded sedimentary facies. The condensed sequences correspond to Arkell's "axes of uplift" (Arkell, 1933) and were later termed "swells", and the intermittent expanded sequences were termed "basins" by Hallam (1958). These areas of basin and swell tectonics are important throughout the Early and Middle Jurassic and in some areas (for example, the Dorset coast) exhibit striking cyclic behaviour, having basinal tendencies at one time and displaying swell tendencies in another (Sellwood & Jenkyns, 1975). It should be stressed that this cyclicity is not synchronous over the whole of the British Isles and other swells are known to have been almost permanent features throughout much of this time, for example, the London-Brabant and Mendip massifs and Market-Weighton swell. Sellwood & Jenkyns (1976) have suggested that both the cyclic and more permanent swell structures were controlled by the underlying faulting in the pre-Permian basement.

The most important phase of regional upwarping and erosion in the British Isles occurred in the early Middle Jurassic (Fig. 1.6b). The Bristol Channel region, formerly a basin in Liassic times, became a slight swell-structure from the Aalenian-Bathonian (Hallam & Sellwood, 1976). Local tectonic uplift occurred at the end of the Lias (post-Toarcian to pre-Aalenian) and in the Bathonian in Yorkshire (Hallam & Sellwood, 1976) and in the Bajocian around the margins of the London-Brabant massif (Hallam, 1975). The Hebrides basin was largely subsiding during the latest Toarcian-Bathonian period, corresponding to a phase of tectonic quiescence in the southern part of the North Atlantic rifting (Masson & Miles, 1986), but during latest Bathonian-Callovian times the region was uplifted during renewed activity (Morton, 1987).

It is this uplift which probably accounts for the regression in northeast England and the extremely shallow Bahamian-type platform development in southern and central England. Updoming in the central and northern North Sea also occurred in the Middle Jurassic (Fig. 1.6b). This is accompanied by the trilete rifting in the region of the Forties-Piper oil field in the Bathonian (Fig. 1.5c). A considerable amount of basaltic fissure volcanicity was associated with the rifting, but waned toward the end of the stage. The presence of montmorillonite-rich clays in the Fullers Earth Clay (Bathonian) of the Bath area, have been interpreted as 'bentonites' or weathered ash-fall

deposits (Hallam & Sellwood, 1976). The clays are too thick to have been deposited from ash-fall from the Forties-Piper volcanic centre and suggest that volcanoes must have been present elsewhere in Britain. The position of these volcanoes is fairly conjectural, it has been suggested that they were upon a landarea such as the London-Brabant Massif (Anderton *et al.*, 1979) or Cornubia (Bradshaw *et al.*, 1992; Fig.1.5c). Bentonites have also been recognised in the Upper Estuarine Series (Bathonian) of the English Midlands and in Skye (Bradshaw, 1975).

Swell structures such as the London-Brabant Massif and the Mendips Swell are known to have fault controlled margins. The basement fault movements are related to the overall tensional regime induced by regional upwarping or as in the case of the North Sea, by diapiric rise of Zechstein salts or 'halokinesis', which is thought to be triggered by crustal movement (Brunstom & Walmsley, 1969). However, it is thought that halokinesis was only important in the North Sea and possibly around Yorkshire (for example the Market-Weighton swell), as elsewhere on the British mainland there is not enough salt (Hallam, 1975). It has also been suggested that the Market-Weighton Swell may have been controlled by an underlying Palaeozoic buoyant granitic intrusion (Bott, 1974). From the Callovian onwards there appears to have been a stabilisation of tectonically controlled basement faults and the British Lias and Dogger swell and basin topography is largely damped out by continuing sedimentation in the Upper Jurassic (Hallam & Sellwood, 1976; Figs. 1.5d and 1.6b). This situation is paralleled in the Mediterranean region, where Tithonian carbonates drape over older block-faulted deposits (Bernoulli & Jenkyns, 1974).

Formation of the Jurassic epeiric sea in north Europe is probably fault controlled and the role of eustatic sea-level rise was probably only a contributing factor (Sellwood, 1972, Sellwood & Jenkyns, 1975). The early rifting in the North Atlantic was probably associated with movement in the Mediterranean region of western Tethys in the Late Lias-Early Dogger (Dewey *et al.*, 1973). This involved a complex system of rifting, with the collapse and subsequent block faulting of an Late Triassic-Early Jurassic carbonate platform, followed by subsidence in Pliensbachian-Toarcian times. This gave way to more subdued rifting in the region in Bathonian times (Bernoulli & Jenkyns, 1974). Widespread subsidence will encourage the formation of marine conditions, and in the central Mediterranean and North Atlantic regions this led to sea-floor spreading and the creation of new oceanic crust (Bernoulli & Jenkyns, 1974). The Forties-Piper rift system and volcanism comprises a "failed-arm" system, which was not later followed by sea-floor spreading.

1.3.3. Palaeoclimate patterns

There is a general consensus that the Mesozoic climate was much more equable than that of the Quaternary, with tropical conditions extending into mid-latitudes and

temperate conditions extending into the polar regions (Colbert, 1964, Frakes, 1979, Hallam, 1975, 1984, 1985). In the Jurassic, subtropical and temperate ferns and gymnosperms extend right up into 60-70° palaeolatitude and there were rich floras in both Grahamland (63°S) and upon the New Siberian Islands (75°N) (Hallam, 1975). That most of these plants appear to have been frost-intolerant, with living relatives that are confined to the tropics, suggests a warm and equable world climate (Barnard, 1973). The wide-ranging distribution of large terrestrial and aquatic reptiles also supports this theory (Colbert, 1964, Hallam, 1975). Temperature-sensitive marine invertebrates, such as reef corals and the giant shelled molluscs which are now confined to the tropics (30°S-30°N), are also found in a wide range of palaeolatitudes in the Jurassic (Arkell, 1956, Hallam, 1975).

There are no Jurassic tillites and associated striated pavements, which are the best indicators of glaciation, and polar ice-caps appear to have been absent in the Mesozoic era. Apart from the palaeontological evidence outlined above, there is very little evidence for palaeotemperatures and temperature distribution in the Jurassic. The presence of thick marine limestones has often been used as evidence to support warm temperatures, and their absence to suggest a cool climate. However, although the former argument is true, the latter is incorrect (Hallam, 1975). Biological carbonate deposition can also be influenced by the amount of terrigenous siliciclastic run-off and volcaniclastic deposition. Therefore, although the principal zone of carbonate deposition in the Middle Jurassic was in the circum-Mediterranean and Arabian peninsula area (a tropical region confined to the western and southern borders of the west Tethys Ocean, Fig. 1.4a; Hallam, 1975), this probably is a reflection of the amount of continental run-off in the other areas and does not suggest a temperature difference between the tropics and the sub-tropics. This is particularly true for the northern European marine deposits and fossil assemblages of the Boreal Realm epicontinental sea (section 1.3.5), which were once thought to represent a cool water environment.

Oxygen isotope determinations of 'absolute' temperatures for the Jurassic have been taken from calcitic fossils, such as belemnites, but they are known to be subject to considerable error (for example, see reviews and discussions in Hallam, 1975, 1985). This is largely the consequence of post-depositional isotopic exchange within the crystal lattice of the calcite, with connate or meteoric waters (Hallam, 1985). The general assumption for the Jurassic palaeotemperatures is that they were fairly warm, and to support ectothermic reptiles (such as crocodiles) and frost-intolerant plants, the high latitudes must have attained at day-time temperature of at least 20°C (Colbert, 1964, Hallam, 1985).

The most useful evidence for Jurassic climate is based upon studies of precipitation or rainfall patterns, as in an equable world the most important seasonal change would be between the wet and dry seasons (Parrish *et al.*, 1982, Hallam, 1984).

The criteria which support a humid environment are: the presence of large coal deposits, indicating the occurrence of large paralic or lacustrine swamps in the Jurassic (Hallam, 1975); bauxites, which indicate a warmer, better drained environment than that for coal deposition (Ronov & Khain, 1962); and chamositic ironstones and kaolinite clay deposition, which indicate continental weathering in warm, humid conditions (Hallam, 1975). The latter two deposits are important as they often occur within shallow marine facies, as well as in terrestrial sequences (Griffin *et al.*, 1968). An abundance of ferns in floral assemblages also indicates a degree of humidity (Hallam, 1984).

Criteria which indicate an arid environment include: aeolian dune sandstones and braided fluvial deposits, indicating a desert environment with flash-floods (Hallam, 1975); evaporites, thick deposits of anhydrite, halite, gypsum and potash salts, signify a low humidity in the air (Kinsman, 1976, Parrish *et al.*, 1982); red-beds and calcrete nodules, which indicate oxidising continental conditions (Hallam, 1984); and an abundance of feldspar in terrigenous sediments, which suggests only a modest amount of chemical weathering (Hallam, 1984).

Plotting the distribution of these criteria on to a palaeogeographic map for the Middle Jurassic, has produced a map of continental humid and arid zones (Fig. 1.4a). Many non-marine sequences do not fall into the convenient humid or arid categories, in displaying features which are characteristic of both regions. For example, in parts of northern Europe and north America , thin coal seams and alternate with evaporites (Hallam, 1975), therefore for these 'mixed' regions a third, intermediate category is proposed. This has been designated "seasonally wet" on Figure 1.4a after Hallam (1985) and indicate a climate of alternating wet and dry seasons. The three precipitation belts for the Middle Jurassic (Fig. 1.4a) are based upon evidence given in Robinson (1971, 1973), Frakes (1979) and Hallam (1975, 1984, 1985; and references therein) and are:

(1) The Humid Belt (Fig. 1.4a). This is roughly a region surrounding the poles at latitudes of about 60º and occurs on or near westward facing northern or southern coastlines, in a belt which would be subjected to strong polar easterly winds (Robinson, 1973). Large deposits of Middle Jurassic coals occur in this belt, particularly on the Siberian Platform, in Japan and in southeastern Australia (Hallam, 1975; Fig. 1.4a). In Japan, there are coarse fluvial sediments which are rich in plant debris. These were most likely deposited within a high-relief, wet climate (Hallam, 1984). There are rich plant beds also upon the Antarctic Peninsula, which indicate a warm humid temperate environment for the continent (Hallam, 1984). The anomalous occurrence of humid conditions in the eastern Asia islands (30ºN), cannot be adequately explained, but there appears to be some confusion as to their exact palaeogeographic position in the Jurassic (Hallam, 1984, 1985).

(2) The Seasonally Wet Belt (Fig. 1.4a). This is roughly a region in mid to high latitudes (30-60º) (Robinson, 1973), including much of northern Europe and the British Isles. The eastern part of Jurassic Pangaea in low and mid latitudes and especially coastal regions surrounding the Tethys ocean are also seasonally wet (Hallam, 1984). These regions would have been subject to moisture-laden monsoonal winds, ensuring a high atmospheric humidity in these regions during the monsoon season (Allen, 1975). In this belt there are Middle Jurassic coal measures in Russia, China and Mongolia, and coals also occur in Mexico, although the rest of the American continent is relatively arid (Hallam, 1984). There are fern-rich plant beds in southern Argentina and alluvial fanglomerates with abundant plant remains, fish, fresh-water mollusca and crustacea indicating a fairly humid environment occur in South Africa (Hallam, 1984). Plant beds of Bajocian age are common in Yorkshire and coals of the same age occur at depth in the Viking Graben. Ironstones and kaolinite clays are a common constituent of Middle Jurassic successions deposited into the European epeiric sea (Hallam, 1984) and in Bathonian terrestrial deposits (Chapter 5). Minor calcretes have also been observed in British Bathonian soil deposits, indicating seasonal or periodic aridity (Chapter 5).

Karstic bauxites are found in the Middle East, although Jurassic coals are rare and most of the Arabian Peninsula appears to have had an arid climate. There are also rare coals in north Africa and on the Indian Peninsula. The Middle Jurassic succession of the Indian subcontinent is mostly shallow marine, but evaporites are of minor significance, which suggests that the climate was seasonally wet (Hallam, 1984).

(3) The Arid Belt (Fig. 1.4a). This occurs in a region surrounding the equator, at low to mid latitudes (0-30º) upon western and central Pangaea. This pattern of precipitation is similar to that of the late Triassic and Lias, proposed by Robinson (1973), and the general aridity of the belt is probably related to the shape and form of the continental masses as a single entity, Pangaea. The most important difference between the Middle Jurassic climate and that of the Quaternary, is the absence of an equatorial humid belt. The shape of Pangaea ensured that zonal winds would have a reduced prominence in the Jurassic (Fig. 1.4), and trade winds would follow the Tethyan coastlines, without passing over large stretches of ocean (Hallam, 1985). These winds would tend to be fairly 'dry' and there would have therefore been a greater dominance of a monsoonal effect around the coastal regions (Parrish *et al.*, 1982). The Laurasian and Gondwanaland promontories would tend to draw the moisture-rich monsoonal winds away from western Pangaea (Parrish *et al.*, 1982).

The arid belt is best defined by the major evaporite-bearing sediments found within most of the central Pangaean area (Hallam, 1975). In the United States Western Interior where a shallow sea had developed in the late Early Jurassic, substantial salt and gypsum deposits are found. These continue into the Gulf of Mexico. Coals are

absent from this region and in South America. Bathonian and Callovian gypsum is worked in northern Argentina, Equador and Peru, and Middle Jurassic continental red-beds are found in Columbia (Hallam, 1984). Most of the Indian subcontinent was marine and fairly humid, during Middle Jurassic time, however, gypsum is found in northwestern India and Pakistan indicating a more dryer environment toward the west. Commercially important Middle Jurassic evaporites have been found in east Africa (Tanzania) and also in Ethiopea and Somalia (Hallam, 1984). Although most of the Middle East is included in the 'seasonally wet belt', the Arabian peninsula had a much more arid climate, which is indicated by the large deposits of commercial anhydrite in the region (Hallam, 1975).

The general patterns of Middle Jurassic global precipitation (Fig. 1.4a) reflect the general palaeogeography of Pangaea. However, these patterns began to break up in the Late Jurassic and Early Cretaceous, with the fragmentation of the supercontinent (see Hallam, 1985, figs.7-8), when the central arid belt grew and then broke up completely by the Late Cretaceous to form climate patterns similar to those of today. The distribution of Mesozoic rainfall patterns can be explained by the changing palaeogeography and increased plate tectonics, however, the general equability of the climate is not so easily explained. Hallam (1985) provides a review and discussion on the four main theories to account for the lack of Mesozoic polar ice-caps, which are : variations in the solar constant; global plate tectonics and continental shift (e.g. Frakes, 1979); variations in the amount of carbon dioxide (CO_2) in the atmosphere, which is also known as the 'Greenhouse Effect' (e.g. Manabe & Wetherland, 1980); and marked deviation between the position of the geomagnetic and geographic poles (Donn, 1982). Hallam (1985) favours an increased amount of atmospheric CO_2 in the Mesozoic which would hinder the formation of polar ice, leading to a more equable climate.

1.3.4. Definition of the Boreal and Tethyan faunal realms

Many of the Jurassic marine invertebrate faunas exhibit an increasing amount of provincialism from the Toarcian and late Pliensbachian stages of the Lias onwards. Arkell (1956) was first to define the global provinces or realms, and he recognised three distinct realms of global ammonite distribution : Tethyan, Boreal and Pacific. However, the "Pacific" realm is no longer in use, as the "Pacific" fauna was not sufficiently distinct from that of the Tethyan (Hallam, 1975). The other realms are defined as :

(1) The Boreal Realm occupied the northern part of the northern hemisphere (Fig. 1.4b), encompassing most of Scandinavia, northern Siberia, Canada, Greenland and a part of northern Europe and the United States (Hallam, 1975). It is best defined by the distribution of certain ammonite families and sub-families. The southern limit of these faunas provides a convenient boundary. However, this boundary

is gradational and shifted to the north and south with time. For instance, in Callovian times the Boreal ammonite fauna spread far into southern Europe and the United States (Fig. 1.4a); whilst in the Bathonian the genera retreated back into northern Europe and Canada, after a late Liassic (Pliensbachian) radiation (Fig. 1.4b) (Hallam, 1975). Other cephalopod groups show a degree of provincialism, however, for most other invertebrates the Boreal realm is marked by the absence of certain Tethyan genera, rather than by the development of indigenous Boreal groups (Hallam, 1975).

(2) The Tethyan Realm covered the rest of the world (Fig. 1.4b) and is marked by a greater diversity of fauna than the Boreal Realm (Anderton *et al.*, 1979). As well as the cephalopods, many groups of invertebrates exhibit this greater diversity, and some groups such as the rudist bivalves, radiolarians, tintinnids and dasyclad algae, are only known from the Tethyan Realm. The faunas are quite cosmopolitan in distribution, however unlike the smaller Boreal Realm, the Tethyan faunas do exhibit a certain amount of provinciality, with smaller faunal "provinces" being distinguishable in the Mediterranean, central Pacific and the coast of east Africa (Hallam, 1975).

There have been many explanations for provinciality in the Jurassic, including physical barriers to migrations, temperature or climatic controls, palaeosalinity controls, and the depth of the sea. Hallam (1975) provides a good review of each of these hypotheses, and suggests a combination of these factors for the restriction of certain groups to the Boreal Realm. The oceanic Tethyan Realm probably had a much more stable palaeoenvironment than the shallow, epicontinental seas of the northern realm. In oceans seasonal fluctuations in climate, salinity and temperature are less noticeable than in the shallow, shelf seas were ecological stresses, seasonal differences and continental runoff are high. Therefore, the low diversity of the Boreal faunas can probably be explained on the more variable and stressful conditions of the epeiric seas (Hallam, 1975).

In the Middle Jurassic, the British Isles occupied a position between the fluctuating boundary of the faunal realms (Fig. 1.4b). For most of the Dogger, southern Britain can be considered to have the closest faunal links with the Tethyan Realm, although ammonites are rare in the Bathonian. In the Callovian the assemblages are much more mixed with the southern spread of Boreal genera (Arkell, 1956; Fig.1.4b). Hallam & Sellwood (1976) have suggested that this may have something to do with the facies changes which occurred from north to south across Europe at that time. Associated with slow eustatic rise in sea-level and tectonic subsidence, quiet water deep marine argillaceous deposits replaced the shallow marine calcareous deposits as far south as the Jura (Hallam & Sellwood, 1976).

In northern Britain the story is more complex, as many formations have seldom yielded ammonites, making correlation extremely difficult. However, the northern British Isles, particularly the Jurassic of Scotland, Yorkshire and the offshore basins

seem to be more closely allied to the Boreal Realm, and therefore, are considered to be 'sub-Boreal'. For instance, distinct Arctic genera have been recovered from as far south as the Viking Graben (Hallam & Sellwood, 1976), which may be explained by the complex tectonic uplift and associated regression to the south of the region barring faunal mixing (Surlyk *et al.*, 1973).

1.3.5. General sedimentation patterns in the British area

In northern Europe the thickness of the Jurassic strata is related to a combination of, and interplay between, two major factors: (a) the uplift and subsequent subsidence of a series of rift-related basins and swells (section 1.3.3) and (b) eustatic changes in sea-level (section 1.2.6) (Hallam & Sellwood, 1976). Sediment thickness variations within the various basins and swells are not related to differential rates of sedimentation, but to the amount of subsidence in these regimes. Relief upon the swells and local land areas was not great, but enough to provide substantial volumes of sediment. The regressive episodes during the late Lias and those which characterise the British Dogger, have produced a complex facies distribution of Middle Jurassic sediments by upsetting the broad equilibrium between subsidence and sedimentation rates which persisted throughout the Lower Jurassic (Sellwood & McKerrow, 1974). Sedimentation rates in the Jurassic are estimated to have been in the region of 1m/2-1m/5 x 10^4 years (Hallam & Sellwood, 1976).

In the British Isles, the Scandinavian and Scottish Shield areas provided the most siliciclastic material to the north (Brand & Hoffman, 1963), although sedimentation patterns in the North Sea and in Yorkshire are concentrically arranged around the Central North Sea High (Bradshaw *et al.*, 1992). Sediment from the Scottish Highlands was fluvially transported into the Midland Valley Graben and Arran Basin. Sediments up to Bathonian age are found on both sides of the Scottish landmass. Upon the Mid North Sea High, Upper Palaeozoic and Triassic sandstones were reworked as siliciclastic debris to the surrounding basins (Hemingway, 1974), including the east Midland shelf and Cleveland Basin in Yorkshire. In the northern North Sea basins, such as the Viking Graben, West Shetland Basin and Unst Basin, the Shetland High provided terrigenous sedimentation (Richards *et al.*, 1988). A similar, but southerly draining alluvial system was apparent in the Central Graben region and over 1000 metres of Aalenian-Bajocian non-marine sediments have been drilled there (Heybroek, 1975). The Forties-Piper lava dome and underlying upwarped Liassic sediments provided much coarse clastic deposition to the basins to the south, north and west. However, the Moray Firth (part of the trilete rift system which formed the volcanic centre) also received a limited amount of marine sedimentation from the west during Dogger times (Bradshaw *et al.*, 1992).

During the early Jurassic, sands were transported eastwards into southern England from early Atlantic rifting, but during the late Lias and mid Jurassic, this sedimentary source was denuded. Hence, in more southern regions the Pennine High, London-Brabant and Welsh massifs may have also contributed a subdued amount of sediment to the east Midland and the Cotswolds-Weald shelf regions and the Wessex Basin. However, the presence of thick sequences of Middle Jurassic carbonates in these shelf-areas which thin progressively landward, suggests that siliciclastic sedimentation was fairly low. In the epeiric marine conditions of central and southern Britain, storm surge ebb currents would be important in redistributing nearshore sediments offshore, as only weak to modest tidal currents are associated with the extensive shelf seas (Metcalf & Underwood, 1995). These storm-deposits and other terrigenous beds interdigitate with the carbonates in regions peripheral to the landareas. In Dorset deep marine deposits occur, indicating a period of subsidence, following a tectonic stillstand in the southwest during the Aalenian-Bajocian.

Lateral and vertical variations in thickness of sediments in the British Isles diminish from the Lower and Middle Jurassic to the Upper Jurassic. This is related to the waning tectonic activity and faulting in the basement (Hallam & Sellwood, 1976). During the Callovian, earlier mid Jurassic depositional basins, became unified by a long period of marine transgression, affecting the whole of the British Isles. The transgression began in the latest Bathonian (*discus* Zone, Fig. 1.3.), depositing the Lower Cornbrash, in what is known as the beginning of the 'Callovian' eustatic cycle (Vail & Todd, 1981, Haq *et al.*, 1987; see below). This sea-level rise continued into the early Callovian, with the deposition of the Upper Cornbrash over most of southern England, the Midlands and into parts of Yorkshire (Callomon, 1955). This led to an improved faunal connection between the Tethyan and Boreal marine realms, in the early Callovian (Callomon, 1955). The transgression continued throughout the early and mid Callovian, with the shallow-water carbonates being onlapped by the deeper, siliciclastic sediments of the Kellaways Beds, followed by the organic rich, deep marine Oxford Clay. Marine conditions spread northwards from southern England and southwards from the Hebridean basin, and the transgression was diachronous over a period of two ammonite zones (*calloviense* Zone-*jason* Zone) (Cope *et al.*, 1980b, Bradshaw *et al.*, 1992).

During the mid part of the stage, the seas transgressed even the structural highs of the earlier Dogger (most notably the Market Weighton 'swell', Bradshaw *et al.*, 1992). In the Cleveland basin shallow marine iron-rich oolites and clays were laid down (Osgoodby Formation). Toward the later part of the Callovian, the anoxic clays of the Lower Oxford Clay, were replaced by much more organic-poor sediments, as the sea became oxygenated once more. Sea-level continued to remain high in the Upper

Jurassic (Fig. 1.3) and replaced the unique terrestrial conditions of the early Dogger with fully marine sediments.

1.3.6. General palaeoenvironment of the British Isles

The British Dogger has a unique series of sedimentary facies, encompassing palaeoenvironments as varied as the deep marine realm to fully terrestrial alluvial plain deposits. This makes it arguably one of the most interesting systems to study in terms of palaeoecological reconstruction. The general palaeoenvironment of the British Dogger is dictated by interplay between global eustatic controls, with the tectonic events caused by the north Atlantic rifting on the shallow epeiric sea of northern Europe.

The Middle Jurassic opened with a development of regressive facies across the British Isles. Fluvio-deltaic environments spread southwards into northern Britain and the North Sea. Whilst, central England and (at times) western Scotland became sites of shallow lagoonal sedimentation. Southern England was generally an area of marine shallows where extensive successions of shallow-water carbonates were laid-down in the absence of terrigenous clastic detritus. In southwest England and further to the south across Europe, fully marine Tethyan conditions prevailed. Periodic transgressions spread these marine conditions northwards. To the north of the Scottish landmass Boreal marine assemblages occasionally transgressed southwards into the region. Normally, however, cephalopods are quite rare in the Dogger and hence, precise stratigraphic and facies correlation is lacking.

The British Dogger climate appears to have been warm and humid and monsoonal storm activity is indicated by shallow marine tempestites (Metcalf & Underwood, 1995) and the presence of charcoal (Chapter 8). The generally low-lying land-areas were lushly vegetated and fringed either by wide alluvial plains or coastal marshes. These in turn provided an excellent environment for preservation of the rich and diverse faunas which they supported. As the shallow marine environment which surrounded the floodplains and marshes was influenced by sluggish tidal activity, most offshore deposition would have been by storm ebb currents.

1.4. Historical and current importance of the Dogger

The Jurassic Period was a time when reptiles dominated vertebrate evolutionary radiation. In the British Lower and Upper Jurassic successions most fossil reptiles sites contain marine forms in association with abundant invertebrates and fish. However, these are supplemented by the important terrestrial reptile finds collected from the subaerial facies of the Dogger (Fig. 1.7). The Middle Jurassic was a key episode in the evolution of terrestrial vertebrates which saw the replacement of primitive groups of amphibians, reptiles and mammals by the modern groups represented today. However,

STAGES	BRITISH ISLES	LAURASIA (Asia, Europe & North America)	GONDWANA (Africa, India, Australia, Antarctica & South America)
TITHONIAN	Purbeck Beds, southern England	Morrison Formation, western U.S.A. Solnhofen Plattenkalk, south Germany	Kirkwood Formation, South Africa
KIMMERIDG.	Kimmeridge Clay, central England & Dorset coast		Tendaguru Beds, Tanzania
OXFORDIAN		Guimarota, Portugal Shangshaximiao Formation, China	Chacarilla Formation, Chile
CALLOVIAN	Oxford Clay, east Midlands	Karatau Mountains, Kazakhstan Xiashaximiao Formation, China	Canadon Asfalto Formation, Argentina Unnamed units, Morocco
BATHONIAN	Forest Marble Formation, central England & Dorset coast Stonesfield & Cotswolds "Slates", Oxon. & Glos. Chipping Norton Formation, Glos.		Walloon Group, Queensland
BAJOCIAN	Great Estuarine Series, Hebrides	Zhenzhunchong Formation, China	
AALENIAN	The Deltaic Series, Yorkshire		Kota Formation, India
TOARCIAN		Navajo Formation, midwest U.S.A. Fengjiahe Formation, China	
PLIENSBACH.		Kayenta Formation, midwest U.S.A.	Aganane Formation, Morocco
SINEMURIAN	Lower Lias, Dorset	La Joya, eastern Mexico Moenave Formation, midwest U.S.A	Botucatu Formation, Brazil Upper Elliot-Clarens Formations, South Africa
HETTANGIAN	Fissure deposits, south Wales & Avon	Newark Supergroup, eastern U.S.A. & Nova Scotia	Forest Sandstone, Zimbabwe

KEY
- Bone-bearing localities
- Footprint localities

Figure 1.7. Terrestrial vertebrate-bearing localities in the Jurassic of the British Isles, Laurasian and Gondwanaland regions.

the Middle Jurassic record of these taxa is poorly known : indeed, until new finds were reported recently from China, and the re-evaluation of the age of the Guimarota lignite fauna of Leira, Portugal from the Upper Jurassic (Kimmeridgian-Oxfordian) to the ?Bathonian-Callovian (Milner and Evans, 1992), virtually the only productive localities of this age anywhere in the world, were in Britain (Fig. 1.7.). Despite the present incomplete record, new British finds are filling crucial gaps in our knowledge of vertebrate evolution.

Historically, the British Dogger has produced some of the most important fossil vertebrate finds. These include the first dinosaur bone to be figured in an academic work (Plot, 1677; Chapter 2). Later discoveries include the first Mesozoic mammals to be described (Buckland, 1828) and the first dinosaur to be named (Buckand, 1824) from the Great Oolite Group (Bathonian) of Oxfordshire (Metcalf & Underwood, in prep., 1994a, b). Following these initial discoveries, and specifically after the popularising of natural science in Victorian England, many large collections of Middle Jurassic fossils were made by dedicated amateurs and professionals. Most provincial museums hold dusty 19th century collections of fossil bones, which have proved the base for modern palaeoecological and biostratigraphic research. Some of the most notable Middle Jurassic fossil collections are those from the Great Oolite Group (Benton & Spencer, 1994; Chapter 2, Appendix A2). Scottish Bathonian vertebrates are well represented in the Hugh Miller Collection (National Museums of Scotland, Edinburgh). Vast collections of Lower Oxford Clay fossils have been made (Martill & Hudson, 1991, their appendix 3), of which the collection of fish from the Christian Malford locality held at Bristol City Museum and Bath Museum and the Leeds Collection of fossil reptiles (which has been broken up and sold to museums around the world) are the most notable.

The mid Jurassic time represented a major evolutionary watershed, when many groups of vertebrates were undergoing diversification and expansion (Figs 1.8 and 1.9; after Benton, 1993a). Diapsid reptiles diversified during the Jurassic into a wide range of new forms, and many Triassic groups continued to radiate during this period (Benton 1990a, d). Archosaurs were the most abundant diapsids on land. It was during the Middle Jurassic that the radiation of many of the 'classic' groups of dinosaurs (large theropods, sauropods, stegosaurs, nodosaurs) took place, replacing the older forms of the Early Jurassic with late Triassic affinities (prosauropods, small theropods, basal ornithischians) (Benton, 1993a; Fig. 1.8b).

Apart from the dinosaurs, crocodilians radiated extensively, and a range of terrestrial carnivorous and piscivorous forms evolved, which are represented in Britain only in the Dogger. The pterosaur groups were undergoing a rapid period of expansion during the mid and late Jurassic (Fig. 1.8b). Lepidosaurs, such as the sphenodontids and the squamates or lizards remained generally small (Fig. 1.8b). Other groups of

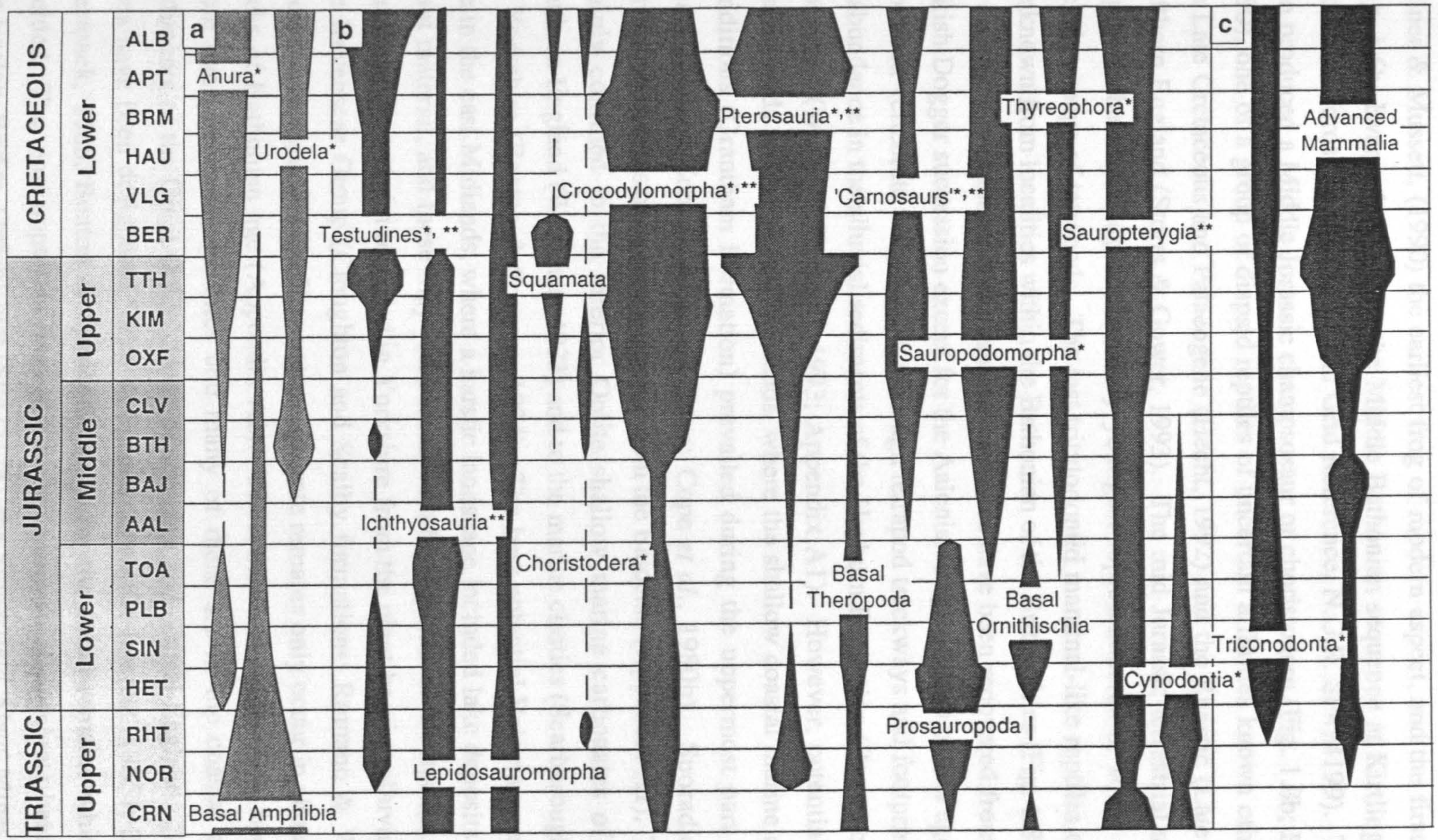


Figure 1.8. The middle Mesozoic fossil record for tetrapods. Groups split into: (a) amphibia; (b) reptilia; c) mammalia. Abbreviations : Stage names: 'CRN' Carnian, 'NOR' Norian, 'RHT' Rhaetian, 'HET' Hettangian, 'SIN' Sinemurian, 'PLB' Pleinsbachian, 'TOA' Toarcian, 'AAL' Aalenian, 'BAJ' Bajocian, 'BTH' Bathonian, 'CLV' Callovian, 'OXF' Oxfordian, 'KIM' Kimmeridgian, 'TTH' Tithonian, 'BER' Berriasian, 'VLG' Valanginian, 'HAU' Hauteririan, 'BRM' Barremian, 'APT' Aptian, 'ALB' Albian; * indicates the presence of members of the group in the Hornsleasow fauna.

terrestrial tetrapods were making notable advances, including frogs (Anura; Fig. 1.8a), salamanders (Urodela; Fig. 1.8a), turtles (Testudines; Fig. 1.8b), and mammals (Fig. 1.8c; Milner, 1993, Benton, 1993a, Stucky & McKenna, 1993). Indeed, Evans, Milner & Mussett, (1988) recently reported the oldest known salamander remains and Evans, Milner & Mussett, (1990) the earliest frog of modern aspect, and the first true lizards (Fig. 1.6b; Evans, 1991) from the Middle Bathonian sequence at Kirtlington Cement Works, Oxfordshire (U.K. National Grid Reference, N.G.R. SP494199). This site has also produced a Middle Jurassic champsosaur or choristodere (Fig. 1.8b; Evans, 1989, 1990), one of a group of diapsid reptiles of uncertain affinities known otherwise from the Late Cretaceous and Palaeogene (Hecht, 1992) and the Rhaetic (Late Triassic) of southern England (Storrs & Gower, 1993). The mid Jurassic terrestrial sites are also important because they contain they youngest representatives of some of the more archaic groups of tetrapods. The last tritylodontid mammal-like reptiles or cynodonts are known from localities within the Bathonian of the British Isles (Fig. 1.8b).

Isolated remains of terrestrial vertebrates have been recovered from most of the British Dogger succession except for the Aalenian stage. Rocks of this age have never produced vertebrate body fossils, although tetrapod trackways and footprints are found in abundance in the alluvial sediments of the Hayburn Formation (Ravenscar Group) in Yorkshire (Whyte & Romano, 1993; Appendix A1). However, potential vertebrate-bearing beds are in the east Midlands, where the shallow coastal marine and lagoonal conditions (Grantham Formation) prevailed during the uppermost part of the stage (*murchisonae* Zone-?*concavum* Zone; Cope *et al.*, 1980b). Sporadic terrestrial vertebrate remains have been recorded from the Bajocian (Appendix A1). The finds are mainly confined to the Inferior Oolite shallow marine carbonates of central and southern England (Waldman, 1973) and to the marine clastics (Scarborough Formation) of Yorkshire (Benton & Spencer, 1994). The best potential Bajocian vertebrate sites are in the east Midlands, where a karstic landscape included lake deposits littered with plant material, and these may yield small vertebrate remains. Footprints of terrestrial vertebrates have been recorded in Yorkshire from the plant-bearing alluvial deposits of the Ravenscar Group (Cloughton and Scalby formations; Romano & Whyte, 1993, Riding & Wright, 1989). However, vertebrate remains only occur in abundance within rocks of Bathonian age (Appendix A2). Terrestrial vertebrates have been recovered from over 50 Bathonian sites and many of these are in the coastal and lagoonal sediments of the Great Oolite Group of southern and central England. Some of these sites have been designated "Sites of Special Scientific Interest" (SSSI; Benton, 1988, Kermack, 1988, Benton & Spencer, 1994), for the fossil remains which they have yielded. The most important English Bathonian terrestrial vertebrate-bearing localities are Enslow Bridge, Oxfordshire (SP475178), the Stonesfield 'Slate' Mines, Oxfordshire (SP387171; SSSI), Woodeaton Quarry, Oxfordshire (SP534122), Shipton-on-Cherwell

Cement Works, Oxfordshire (SP475178; SSSI), Sharps Hill Quarry, Oxfordshire (SP338358) Huntsmans and Brockhill Quarries, Gloucestershire (SP1225; SSSI, and SP133236), New Park Quarry, Gloucestershire (SP171296; SSSI), Hornsleasow Quarry, Gloucestershire (SP1313220), Tarlton Clay Pit, Gloucestersire (SO970001), Watton Cliff, Dorset (SY451908-452907; SSSI), Swyre, Dorset (SY525868), Leigh Delamere petrol station section, Wiltshire (ST890790) and the Atford area, Wiltshire (ST8666). These all occur in the Great Oolite succession. Another important terrestrial vertebrate locality is Loch Scavaig, Skye (NG531164, NG 519154; SSSI; pers. obs., 1993), which occurs in the lagoonal and shallow marine facies of the Great Estuarine Group. Vertebrate finds have been made on the Isle of Eigg (Miller, 1858, Hudson, 1966; pers. obs., 1993). Although, numerous dinosaur footprint and trackway horizons are recognised in the Scalby Formation of north Yorkshire, no body fossils have been recovered (Riding & Wright, 1989; pers. obs., 1993). Similar swampy sediments occur in the Rutland Formation of the east Midlands, and rare vertebrate remains have been found there (Benton & Spencer, 1994). Further sites and additional stratigraphic, palaeontological and palaeoenvironmental information on each locality are given in Appendices A1-A2. References include Woodward (1894), Kermack (1988), Benton (1988), Benton & Spencer (1995), Metcalf & Underwood (in prep. 1994a, b) and Evans & Milner (1994).

In the marine realm, reptiles were also undergoing a period of diversification during the early and mid Jurassic. Ichthyosaurs and plesiosaurs (sauropterygians) are extremely abundant and diverse within British Jurassic successions (Benton & Spencer, 1994). However, except for within the Lower Oxford Clay (Callovian; Appendix A3), their remains are not nearly so abundant or well-preserved in the shallow marine facies of the Dogger as they are within the marls and clays of the Lias and Malm. Marine crocodiles such as the piscivorous teleosaurs and steneosaurs are well represented in the Dogger, and they too were undergoing a rapid period of evolutionary diversification (Fig. 1.8b). The superbly aquatically-adapted metriorhynchid crocodiles or geosaurs are first known from the Lower Oxford Clay localities near Peterborough, Cambridgeshire and from localities in Buckinghamshire (Martill & Hudson, 1991).

Fishes underwent a significant period of change within the Middle Jurassic (Fig. 1.9; after Cappetta *et al.*, 1993, Gardiner, 1993 and Patterson, 1993) and many of the shallow marine sediments of the British Dogger contain abundant disassociated remains of cartilaginous fish (chondrichthyians; Fig. 1.9a-c) and bony fish (osteichthyians; Fig. 1.9d, e). Whole fish skeletons are rare in the Dogger, although they occur within the Lower Oxford Clay. Important changes in the evolutionary record of the chondrichthyes are the appearance and diversification of many of the neoselachian groups (Fig. 1.9b), which eventually replaced the more archaic selachians such as the hybodonts and acrodonts in the Mesozoic (Fig. 1.9a) (Cappetta, 1987,

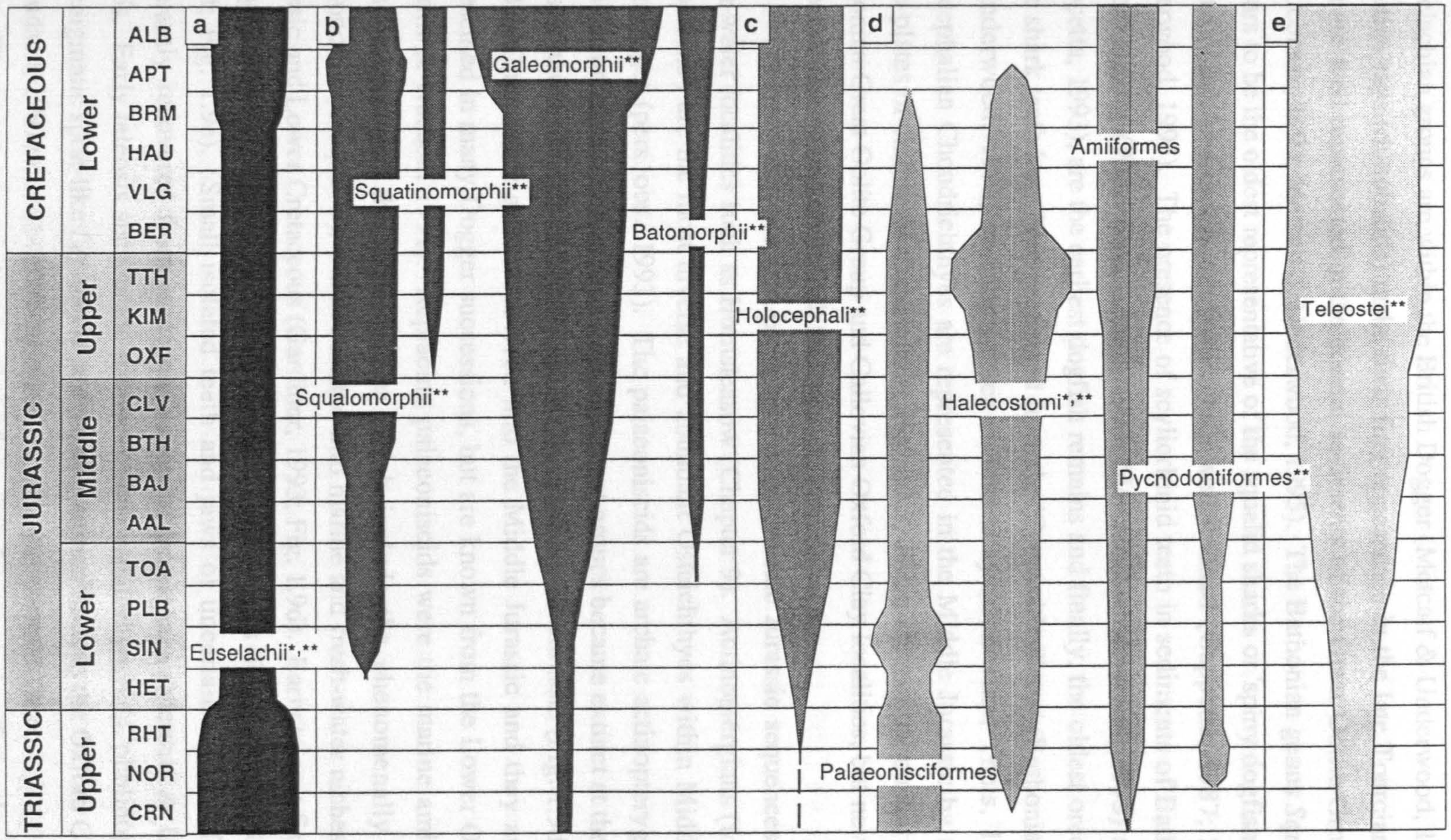


Figure 1.9. The middle Mesozoic fossil record for fish. Groups split into (a) euselachii or old sharks; (b) neoselachians; (c) holocephalians, including chimaeroids; (d) holosteans; (e) teleostei. For abbreviations and explanations see Fig. 1.8.

Cappetta *et al.*, 1993). The neoselachians include the galeomorphs (great sharks; Fig. 1.9b), the batomorphs (rays and skates; Fig. 1.9b), the squalomorphs (lesser sharks; Fig. 1.9b) and the squatinomorphs (monk fishes; Fig. 1.9b) (Steel, 1985) and are the most successful living group of Chondrichthyes. Some of the first occurrences of neoselachian groups are within the British Dogger (Metcalf & Underwood, 1995). The primitive rays (rhinobatids) make their first appearance in the late Toarcian and their teeth are well represented in the coastal sediments of the Great Oolite Group (Ward, pers. comm., 1993; Metcalf & Underwood, 1995). The Bathonian genus *Squalogaleus* appears to be the oldest representative of the squalid sharks or 'spiny dogfish', although its taxonomic position is not particularly well defined (Cappetta, 1987; Metcalf & Underwood, 1995). The presence of scyliorhinid teeth in sediments of Bathonian age in both Britain (Ward pers. comm., 1993; Metcalf & Underwood, 1995) and Africa (Cappetta, 1993), are the earliest dogfish remains and finally, the oldest orectolobid or nurse shark teeth have been recovered from the 'Cotswolds Slates' (Bathonian; Metcalf & Underwood, 1995) and the Lower Oxford Clay (Callovian; Theis, 1983). The holocephalian Chondrichthyes are represented in the Middle Jurassic by spines and tooth-plates of chimaeras ('rabbit fish'; Fig. 1.9c). These are most abundant within the Bathonian Great Oolite Group and Callovian Oxford Clay localities, but never form an important component of the fish faunas.

The bony fish are also present in many Middle Jurassic sequences, including fresh-water localities such as Hornsleasow (Chapter 9). Actinoptergians (formerly the 'holosteans') are the most diverse and abundant Osteichthyes within Middle Jurassic assemblages (pers. obs., 1993). The palaeoniscids are archaic actinopterygians which were dominant during the late Palaeozoic, most groups became extinct at the end of the Triassic and during the Pliensbachian Stage of the Lower Jurassic (Fig. 1.9d; Gardiner, 1993). Only a few genera survived into the Middle Jurassic and they are not well represented in many Dogger successions, but are known from the Lower Oxford Clay (Martill & Hudson, 1991). Replacing palaeoniscids were the marine amiiforms and pycnodontids, and the halecostomes which include the phenomenally successful semionotids (Chapter 9), which radiated into marine and fresh-water niches during the Jurassic and Lower Cretaceous (Gardiner, 1993; Fig. 1.9d). Early teleost fish are also a fairly common find in Middle Jurassic marine assemblages (Patterson, 1993, pers. obs., 1994; Fig. 1.9e). Small isolated teeth and jaws of uncertain teleost affinities are frequently recovered from acid preparation of limestones (Metcalf & Underwood, 1995). Early teleosts such as the pachycormids, aspidorhynchids, pholidophorids and the enigmatic sprat-like *Leptolepis* are not uncommon finds in the Oxford Clay (Martill & Hudson, 1991).

The most spectacular remains of marine faunas derive from the bituminous shale units of the Lower Oxford Clay, and several important museum collections of

Oxford Clay vertebrates have been made (for review, see Martill & Hudson, 1991). These remains are often complete, or nearly complete, articulated skeletons, the result of their original deposition on undisturbed stagnant bottom waters unique to the northern European Jurassic shelf sea (Martill & Hudson, 1991). The marine vertebrates from the Oxford Clay are particularly well preserved and form a centre-point of all international taxonomic studies. Indeed, the Lower Oxford Clay locality at Christian Malford, Wiltshire (ST957774), may be the only British fossil Lagerstätte. This site yielded extremely large numbers of beautifully articulated fish and some reptiles in the 19th century (Egerton, 1843), but sadly they were collected from a temporary exposure and further collections cannot be made (Martill & Hudson, 1991).

The most important marine vertebrate-bearing localities in the Dogger are Leckhampton Quarries, Gloucestershire (SO950185) and Crewbourne, Dorset (ST458104) from the Inferior Oolite (Bajocian), Stonesfield, Oxfordshire (SP387171), Huntsmans and Brockhill Quarries, Gloucestershire (SP1225 and SP133236), Islip Ironstone Quarry, Northamptonshire (SP975782), Norman Cross Brickworks, Cambridgeshire (TL170912), and Kildonnan and Eilean Thuilm, Eigg (NM495870 and NM483913; SSSI), which are all of Bathonian age. The Callovian sites include Backwater Beach, Weymouth, Dorset (SY677790), Putton Lane Brick Pit, Dorset (SY650801), Chippenham, Wiltshire (ST9173), Summertown Brick Pit, Oxfordshire (SP5109), Wolvercote Brickpit, Oxfordshire (SP494105), Shellingford Crossroads Quarry, Oxfordshire (SU326942), Stewartby Clay Pit, Bedfordshire (TL0142), Eynesbury Brick and Tile Works, Cambridgeshire (TL1859), St. Ives Brickyard, Cambridgeshire (TL304718), Woodston Lodge, Cambridgeshire (TL1897) and Fletton Brick Works, Cambridgeshire (TL1995). Sporadic discoveries have been made in the Cornbrash and Kellaways Beds. Additional information about these and other aquatic vertebrate-bearing localities is given in Appendices A2-A3. References include Phillips (1871), Woodward (1895), Arkell (1933), Leeds (1936), Martill (1986, 1988), Martill & Hudson (1991), Metcalf & Underwood (1995) and Benton & Spencer (1995).

Although the reptile and fish remains recovered from the Lower Oxford Clay sites are often exceptionally preserved and fairly common, no locality has been selected by English Nature as an SSSI. Benton & Spencer (1995) consider this appropriate as many of the older sites are lost, infilled or degraded, and active pits are often "worked in a way that prevents the protection of fossiliferous horizons" (Benton & Spencer, 1995, p.119). These faunas have been described elsewhere (Martill & Hudson, 1991, and references therein) and a review will not be attempted here. The vertebrate faunas of the Oxford Clay represent what were mainly pelagic ecological niches, and they shall be used for comparative purposes only in this thesis. Vertebrate remains and assemblages continue to be discovered throughout the British Dogger succession, particularly in actively worked quarries and eroding cliffs. For instance in the past year

(1993-1994) terrestrial microvertebrates have been recovered from a new White Limestone site at Merton, Oxfordshire (P. Powell, pers. comm. 1994), and a previously uncatalogued 'Cotswolds Slate' locality has been discovered at Daglingworth Quarry, Gloucestershire (N. Hollingworth, pers. comm. 1994) (Appendix A2). Also sampling methods for vertebrates are changing toward a much more comprehensive search for all fossil remains - i.e. NOT just the beautifully preserved, the articulated, the large or the unusual. Hence, old sites are being resampled for new types of vertebrate remains, and the acid preparation of many bone-bearing Bathonian limestones has yielded a suite of microvertebrate material. This is the case with the author's re-examination of the 'Cotswolds Slates' quarries around Eyford Hill and Sevenhampton Common in the northern Cotswolds (the Cotswolds Slates have now been formalised as the Eyford Member of the Charlbury Formation, Boneham & Wyatt, 1993, Metcalf & Underwood, 1995). These shallow-marine calcareous tempestites produced a similar, but less diverse, fauna to the more famous 'Stonesfield Slates' of Oxfordshire (now part of the Taynton Formation, Boneham & Wyatt, 1993) in the nineteenth and early twentieth centuries (Savage, 1963). Re-sampling and acid preparation of the arenaceous limestones has revealed a wealth of marine microvertebrates not previously described from the British Bathonian (Metcalf & Underwood, 1995). In short, the state of Middle Jurassic vertebrate research is extremely healthy at the present.

1.5. Vertebrate-bearing coastal-plain and marginal marine environments in the Dogger

1.5.1. The fluvio-deltaic environment of the northern British Isles

The marine environments of the Lias gave way over much of northern Britain to fluvio-deltaic sedimentation in the late Aalenian, following domal upwarp in the North Sea and a stillstand in eustatic sea-level. The central and northern North Sea, east and west Scotland and the Cleveland Basin of Yorkshire were the main areas of fluvio-deltaic deposition in the British Dogger. In Yorkshire there were four episodes of alluvial facies deposition upon a broad coastal plain, namely the Hayburn or Saltwick Formation (formerly the Lower Deltaic Series; Aalenian, Cope *et al.*, 1980a), the Gristhorpe and Scycarham Members of the Cloughton Formation (formerly the Middle Deltaic Series; Lower Bajocian, Cope *et al.*, 1980a) and the Scalby Formation (formerly the Upper Deltaic Series; Upper Bajocian-Bathonian, Cope *et al.*, 1980a). These were separated by interludes of transgressive marine deposition and have been placed together in the Ravenscar Group sequence (Hemingway & Knox, 1973, Cope *et al.*, 1980a).

The river channels of the upper Hayburn Formation (*murchisonae* Zone-*concavum* Zone) are seen to cut down into the underlying marine strata of the Early Aalenian (Dogger Formation; *opalinum* Zone) and this suggests that there may have

been a significant fall in sea-level toward the end of the regressive phase (Livera & Leeder, 1981). A typical Ravenscar Group cyclic succession would be an abrupt lithological change from coarsening-upwards siliciclastic marine deposits or carbonates into non-marine clastics (Hemingway, 1974). The coarsening-upwards sequence appear to represent shoreline deposits, which are cut and eroded by the fining-upwards river deposits. The overall transport of sediment appears to have been from the northwest toward the southeast, and appear to be derived from the Upper Palaeozoic and Triassic of the Mid North Sea High (Sellwood & Hallam, 1974, Nami, 1976). There is also evidence of drainage from the west (?Pennine High) by the low sinuosity channels of the Hayburn Formation (Hancock & Fisher, 1981, Livera & Leeder, 1981). In the northern part of the Cleveland Basin, freshwater lakes surrounded by a horsetail-dominant flora (Hemingway, 1974), while to the southeast the rivers drained into brackish swamps drained by tidally influenced channels (Hancock & Fisher, 1981). The alluvial and deltaic conditions prevailed throughout the upper Aalenian into the Early Bajocian (Cloughton Formation; *?discites* Zone-*sauzei* Zone). However, during the latest part of the Early Bajocian (*sauzei* Zone), three pulses of marine transgression occurred within the basin, from the south and the east, and the marine Scarborough Formation shales were deposited (Bate, 1965, 1967). Following the marine lowstand in the latter part of the stage (*subfurcatum* Zone), an alluvial environment was established in northeast England and much of central England. At this time rivers draining from the north and northwest off the Mid North Sea High, deposited the coarse braided channel alluvium known as the Moor Grit (Scalby Formation; Leeder & Nami, 1979, Fisher & Hancock, 1985). Non-marine sedimentation continued into the Bathonian, although the exact dating of the Scalby Formation is uncertain (Fisher & Hancock, 1985). Stratigraphic evidence suggests that the thick non-marine depositional units represent relatively short periods of time in comparison with the intervening marine formations (Hallam & Sellwood, 1976, Lott & Humphreys, 1992). As sea-level rose in the Late Bathonian, low-sinuosity braided streams were replaced by meander belts upon the coastal plain (Leeder & Nami, 1979).

The fluvial cyclothems are frequently associated with rootlet beds, plant debris, charcoal fragments and thin coals. Locally remains of the horsetail *Equisetites* are preserved in upright positions, indicating inundation during rapid flooding events. The floral assemblages include ferns, horsetails, ginkgoales, conifers, and bennettitalean gymnosperms. Spectacular meander belts and palaeochannels are also exposed locally (e.g. Nami, 1976, Nami & Leeder, 1978). Dinosaur and other tetrapod footprints are common in all the non-marine formations (Sargeant, 1974, Delair & Sargeant, 1985). They commonly occur as partial trackways upon dried-up fluvial overbank muds and sandstone stringers (e.g. Whyte & Romano, 1992). Some of the beds in the Scalby Formation such as the Footprint Bed (Sargeant, 1974; Appendix A2) and beds exposed

at White Nab (pers. obs.; Appendix A2) are extremely churned-up by what has been termed 'dinoturbation' (Lockley, 1991), making it impossible to trace an individual trackway over a great distance. However, fossil bone is rare in these deposits and only a few pieces of marine reptile and fish have been recovered from the Scarborough Formation (marine; Bajocian) at White Nab (Fox-Strangways, 1903; Appendix A1). The lack of bone within the non-marine formations of the Ravenscar Group must be dictated by preservational and possibly sampling factors, the soils appear to have been largely acidic and rather swampy in character (Hemingway, 1974) which would have hindered preservation. No bone moulds have been discovered in the fine-grained sands and silts (pers. obs., 1993), which suggests that acid dissolution of bone material was an early diagenetic feature and that moulds may have been subsequently destroyed by compaction. However, sampling oversight cannot be eliminated, as abundant moulds of unionid freshwater bivalves are found in some localities (Anderton *et al.*, 1979).

In east Scotland, a fluvio-deltaic succession also developed near Brora and offshore in the Moray Firth Basin (Late Aalenian-Bajocian; Linlay *et al.*, 1980). Freshwater sands are overlain by a sequence of deltaic muds and coals with fresh- to brackish-water assemblages. A thick coal seam at the top of the succession suggests that fluvial abandonment, led to an accumulation of a thick peaty deposit in a coastal swamp (Barnard & Cooper, 1981). Similar Middle Jurassic fluvio-deltaic sequences are also known from the offshore Brent oil field region (Rannoch and Etive formations, Late Aalenian, Ness Formation, Late Aalenian-Bajocian; Bradshaw *et al*, 1992) but these are never likely to yield vertebrate material.

In northwest Scotland, siliciclastic sedimentation into the Hebrides Basin was largely from the Scottish Highlands and Hebrides Platform throughout the mid Jurassic. The Aalenian-Bajocian of the Hebrides Basin is represented by 200-485 metres of micaceous marine sandstones and coarse-grained quartzites, known as the Bearreraig Sandstones (Upper Toarcian-Upper Bajocian; Morton, 1965, Hudson, 1983). The micaceous sandstones coarsen upwards into the quartzites and probably represent an accretionary shoreface environment (Morton, 1965). The Bearreraig Sandstone Formation is topped by the Garantiana Clay (*garantiana* Zone), a thin dark band of clay packed full of ammonites (Hudson, 1983), which passes vertically into the basal oil shales of the Cullaidh Shale Formation - the basal unit of the marginal marine deposits Bathonian Great Estuarine Group (Hudson, 1983). Fish remains are not uncommon in most of the formations (Appendix A2; pers. obs., 1993), but tetrapods have only been recovered from the Kildonnan Member (Lealt Shale Formation; ?Lower Bathonian) of Eigg (Hudson, 1966) and the Kilmaluag Formation (Middle-Upper Bathonian) of Skye (Waldman & Savage, 1972). A single vertebrate footprint, thought to be from a theropod dinosaur, was also recovered from a cliff-section of Lealt Shales (Lonfearn Member) in Skye (Andrews & Hudson, 1984). There is much potential for further

vertebrate finds in the Great Estuarine Group, especially in rocks which have undergone little or no contact metamorphism from the overlying and intrusive Tertiary igneous complex. These lagoonal and shallow-marine sediments are overlain in the latest Bathonian by the mottled, alluvial clays of the Skudiburgh Formation (Hudson, 1983). The Scalby Formation and Great Estuarine Group have not been correlated with the standard ammonite zonal scheme employed in the southern British Isles, although many localised schemes have been proposed based upon a variety of other macro- and microfossils (for review, see Cope *et al.*, 1980a).

1.5.2. The marginal marine environments of central England

Aalenian strata in southern England and the Cotswolds are described as the Lower Inferior Oolite, it is lithologically similar to the Bajocian Middle and Upper Inferior Oolite, and the subdivision is made upon faunal content. In the English Midlands, modest uplift in the early Aalenian led to the erosion of late Liassic sediments and deposition of a shoreface facies (Northampton Formation) against the northern edge of the London-Brabant massif, in an area known as the East Midland Shelf (Bradshaw *et al.*, 1992). The overlying non-marine silts of the Grantham Formation, are more easily separated from the overlying Bajocian carbonates in the English Midlands.

Bajocian sedimentary deposition in the British Isles occurred within the same depositional basins as the underlying Aalenian. Marine transgression from the south onto the east Midland coastal-plain, initiated the deposition of the Lincolnshire Limestone Formation, in the earliest Bajocian (*discites* Zone; Bradshaw *et al.*, 1992). During the late Lower Bajocian (*laeviuscula* Zone-*sauzei* Zone) the shelf was later exposed during a period of sea-level stillstand and a facies change to non-marine sedimentation occurred marking the progression of the paralic deltaic and lagoonal complex spread southwards from the Cleveland Basin. A karstic terrain developed upon the exposed early Bajocian limestones. In southern and central England, the extremely fossiliferous Middle and Upper Inferior Oolite were laid down upon the Cotswolds-Weald Shelf and passed into open marine carbonate deposition further to the south.

Throughout the mid Jurassic the Cotswold-Weald shelf was subject to tectonic and eustatic controls and developed a sequence of marginal marine facies which are quite unique in the British Dogger. Regressions were usually marked by the progradation of siliciclastic sediments with stronger freshwater influence, whereas transgressions resulted in the eastward and northeastward spread of shelf carbonates, in the form of offshore oolite shoals (Metcalf & Underwood, in prep., 1994a).

The Bathonian rocks of England are often referred to as the "Great Oolite Series" (Torrens *in* Cope *et al.*, 1980b). The Great Oolite Group (*sensu stricto*) encompasses the Lower Fullers Earth (and its lateral variations; *zigzag* Zone-?*hodsoni*

Zone), the 'Great Oolite' (*sensu* 19th-early 20th century workers e.g. Woodward, 1894, Richardson, 1911, 1929; *zigzag* Zone-*aspidoides* Zone), the Forest Marble Formation (*aspidoides* Zone-*discus* Zone) and the Lower Cornbrash (*discus* Subzone) (Cave, 1977). The shallow-marine and coastal sediments of the Great Oolite Group were deposited upon the Cotswolds Weald Shelf, in central and southern England. The Lower Fullers Earth deep marine clays and the shallow marine carbonates of the 'Great Oolite' interdigitate in many sections, although the Fullers Earth is mainly confined to a line south of Cirencester and the Cotswolds hills. Swamps covered the east Midland Shelf in the Lower Bathonian (*?tenuiplicatus* Zone-*progracilis* Zone; Bradshaw *et al.*, 1992), and a rhythmic succession of non-marine and marine rocks were laid down upon the shelf during the rest of the stage (Bradshaw, 1978). These deposits were once known as the "Upper Estuarine Series", but are now formalised as the Rutland Formation (Hallam, 1992).

The Great Oolite Group of the Cotswolds is a classic example of this interplay between non-marine and marine influences. The London-Brabant and possibly the Pennine landmasses provided terrigenous clastic material at times of low sea-level stands. Marine incursions into the lagoonal belt allowed oyster (usually *Praeexogyra*) beds to be established (Hudson & Palmer, 1976), while at other times, rootletted beds and variegated clays, interpreted as coastal marsh soils are found in association with tetrapod bones, freshwater gastropods (*Bathonella* or *Viviparus*) and ostracodes, unionid bivalves and charophytes (e.g. Sellwood & McKerrow, 1974, Palmer, 1972, 1979, Metcalf *et al.*, 1992, Evans & Milner, 1994). Lagoonal facies were occasionally exposed (Palmer & Jenkyns, 1975) and the carbonate shelf itself was periodically exposed to karstification and terrestrialisation (Chapters 4-6). There is a complete gradation between facies containing restricted and euryhaline invertebrate communities to those with more normal marine affinities (Sellwood, 1978). Corals, terebratulids and echinoids inhabit the least restricted lagoons (Metcalf & Underwood, in prep., 1994a) and cephalopods are quite rare. Salinity variation and strong seasonality were probably the important faunal-inhibiting factors within these lagoons. Terrestrial vertebrate faunas from a case study within the Great Oolite Group of the Cotswolds form the basis for the succeeding chapters of this thesis and will not be described in any detail here.

1.6. Scope and aims of the research

The general regressive nature of the Dogger system in the British Isles, has provided an almost unique opportunity to study Middle Jurassic vertebrate faunas within terrestrial and shallow-marine, coastal palaeoenvironments. In the first chapter I have outlined the general state of research on what is known about the environment of the British and northern European Dogger. The scope of this thesis is then to provide a review of the sedimentary and preservational (that is 'taphonomic') environment of Middle Jurassic vertebrate-bearing horizons.

In Chapter 2 and Part B (Chapters 3-6), a case history of a vertebrate-bearing locality in the British Bathonian sequence is described and its depositional conditions are studied and contrasted with other Middle Jurassic localities. Chapter 2 introduces the locality and the original discovery and sampling of the fauna. Chapters 3-6 provides evidence that the case study represents a fully terrestrial faunal assemblage in a coastal-plain sedimentary regime. In Chapter 7 the concept of vertebrate taphonomy and in particular those parameters applied to terrestrial and aquatic preservation of vertebrate remains are explained. Also reviewed are the principles of invertebrate taphonomy. These factors are then considered for the non-vertebrate material (Chapter 8) and the vertebrate material (Chapters 9-10) separately.

In the final Part C (Chapter 11) vertebrate taphonomy and palaeoecology is reviewed. There will also be emphasis put upon the problems faced by working with extinct biotas and incomplete faunal assemblages in trying to build up palaeocommunities and ecological niches. The concluding remarks include directions for future research, and suggestions for resolving these problems.

In summary the specific aims of this work are :

1) To combine depositional environmental studies with preservational factors for coastal-plain assemblages;
2) To review the state of vertebrate taphonomy for non-mammalian microvertebrate assemblages;
3) To review the palaeoecology of the British Middle Jurassic vertebrate palaeocommunities.

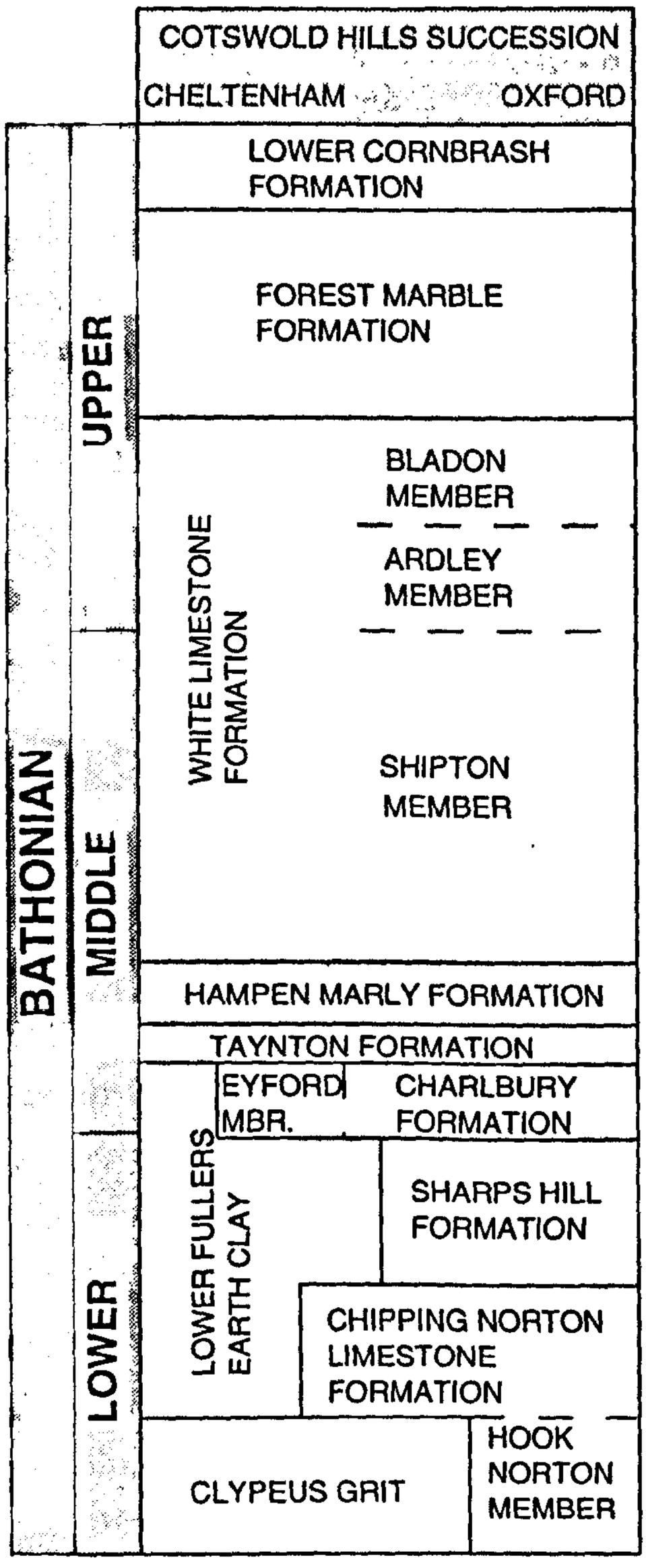
Chapter Two Hornsleasow: an introduction

2.1. Why study the Hornsleasow fauna?

The Great Oolite Group (Bathonian, Middle Jurassic; Fig. 2.1) of the Cotswold hills is a thick succession of nearshore marine, lagoonal and coastal deposits. These sediments were among the first deposits to be systematically collected for Mesozoic vertebrate fossils. They are of international significance as sources of Middle Jurassic vertebrates, particularly terrestrial tetrapod fossils, which have a poor fossil record elsewhere in the world (Chapter 1, Fig. 1.7; Benton 1988, Evans & Milner, 1993, Metcalf *et al.*, 1992). Some of the most important Great Oolite Group terrestrial vertebrate sites are shown in Fig. 2.1 and, in all, over forty localities have yielded vertebrate remains (Benton & Spencer, 1995; Appendix A2). The majority of the vertebrate material was found in the 19th and early 20th century, and the most common finds are the large disarticulated remains of reptiles. Many provincial museums are littered with such material, some of which is inadequately labelled and often in need of attention. However, there are some excellent collections which include type specimens, for example the "Stonesfield Slate" collections at Oxford University Museum (Appendix A2) and "Cotswolds Slates" fossils housed at Gloucester and Bristol City Museums (Savage, 1963, Metcalf & Underwood, 1995).

Unfortunately, more than half of the sites are disused or overgrown quarries, trial pits and railway sections; and some have been infilled or the location forgotten. However, the Great Oolite Group lithologies continue to be worked in a number of quarries (Fig. 2.1 and Appendix A2 list the working outfits). Also five fossil reptile localities are protected as "Sites of Special Scientific Interest" by English Nature (Benton, 1988, Benton & Spencer, 1995; Appendix A2). In more recent excavations, both large vertebrate remains ("macrovertebrate" material) and small vertebrate remains ("microvertebrate" material - usually elements under 10 millimetres) are sought and recovered.

The sampling of a recently discovered Lower Bathonian (166 million years ago; Harland *et al.*, 1991) vertebrate site at Hornsleasow in northern Gloucestershire (Fig. 2.2) has revealed a wealth of vertebrate fossils, small and large. The Hornsleasow site is of especial interest in the context of the Cotswolds Middle Jurassic since it lies at a significantly earlier stratigraphic level than the other British Bathonian vertebrate-bearing localities. As the Hornsleasow fauna contains early examples of frogs, salamanders, pterosaurs and other small forms, otherwise known only from the Upper Jurassic (such as the Morrison Formation dinosaur localities of Montana, Wyoming,



Tarlton Clay Pit, Glos.^	freshwater lagoon
Enslow Bridge Quarry, Oxon.	freshwater lagoon
Kirtlington Quarry, Oxon.^	freshwater lagoon
Woodeaton Quarry, Oxon.^	freshwater lagoon
Sarsden Quarry, Oxon.	freshwater lagoon
Sharps Hill Quarry, Oxon.	freshwater lagoon
Temple Mills Quarry, Oxon.	freshwater lagoon
New Park Quarry, Glos.^	?karst
Hornsleasow Quarry, Glos.^	**freshwater karst**
Smith's Quarry, Oxon.^	?karst

^ Extant section/active quarry (see Appendix A)

Figure 2.1. Freshwater and terrestrial vertebrate-bearing localities in the Great Oolite Group (Bathonian) succession of the Cotswolds. Glos. = Gloucestershire; Oxon. = Oxfordshire.

and Utah; Kimmeridgian-Tithonian, Dodson, *et al.* 1980) and the Bathonian site at Kirtlington, Oxfordshire, it was realised that some of these may be the oldest representatives of their respective groups (Metcalf *et al.*, 1992). The replacement of the early Jurassic faunas during a major episode of terrestrial tetrapod evolution in the early-middle Jurassic is not well known, and therefore the Hornsleasow assemblage may have phylogenetic significance in establishing basal characters within these newly arisen groups (Benton *et al.*, 1995).

The terrestrial nature of the site is also of interest, as sedimentological and palynofloral interpretations imply that the vertebrate material is *in situ*, suggesting that there was a significant retreat of marine conditions off the west Midlands shelf, during the earliest Bathonian, possibly associated with renewed tectonic uplift on Triassic basement block-faulting.

2.2. History of previous research at Hornsleasow and current status of knowledge

The Chipping Norton Formation has been associated with the recovery of dinosaur bones from the early 19th century onwards; but in fact the first recorded specimen of a dinosaur bone was discovered in 1677 at Cornwell, near Chipping Norton, Oxfordshire from the upper part of the Inferior Oolite succession (Plot, 1677), and was most probably recovered from the Chipping Norton Formation. Much disarticulated reptile material has since been recorded from this formation, including specimens of the sauropod dinosaur *Cetiosaurus* (Beesley, 1877; Phillips, 1878; Richardson, 1911), the theropod *Megalosaurus* (Richardson, 1911, 1929) and the stegosaur *Lexovisaurus vetustus* (Galton & Powell, 1983). Bathonian reptile sites have been fully documented by Benton (1988), Evans & Milner (1994), Benton and Spencer (1995) and a list is given in Appendix A2.

The section at Hornsleasow Quarry has been described and documented by several stratigraphers, most notably Richardson (1929), Channon (1950), Torrens (1968, 1969), Vaughan (1989) and Metcalf *et al.* (1992). The most recent work has been done upon the excavation and succeeding palaeoenvironment project upon the Hornsleasow clay horizon, which is documented here. This work has also been described in a series of unpublished reports for Gloucester City Museum and several papers, by Cole (1989), Darlington (1989), Vaughan (1988, 1989), Metcalf *et al.* (1992), Metcalf & Walker (1994), and Metcalf (1994).

2.3. Location and extent

Hornsleasow Quarry (or Snowshill Quarry as it was known in the earlier literature) is located in the northern Cotswolds hills, some 5km northwest of the market town Stow-in-the-Wold, and to the east of Gloucester and Cheltenham (U.K. National Grid Reference SP131322; Figs. 2.2b and 2.3). It is owned by Huntsman's Quarries Limited

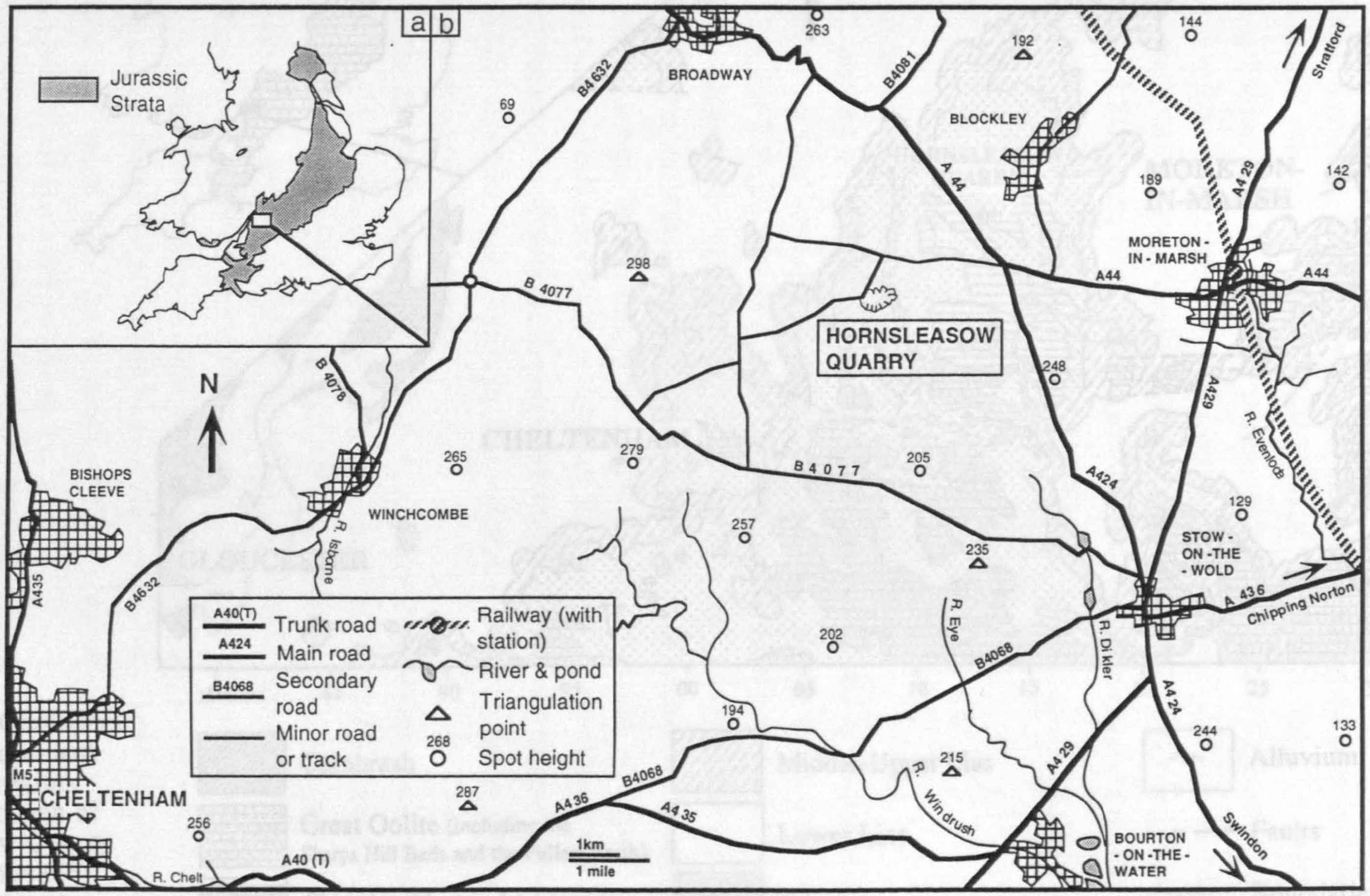


Figure 2.2. The extent of Jurassic outcrop (a) in southern Britain and (b) location of Hornsleasow Quarry in northern Cotswolds. North (N) is indicated.

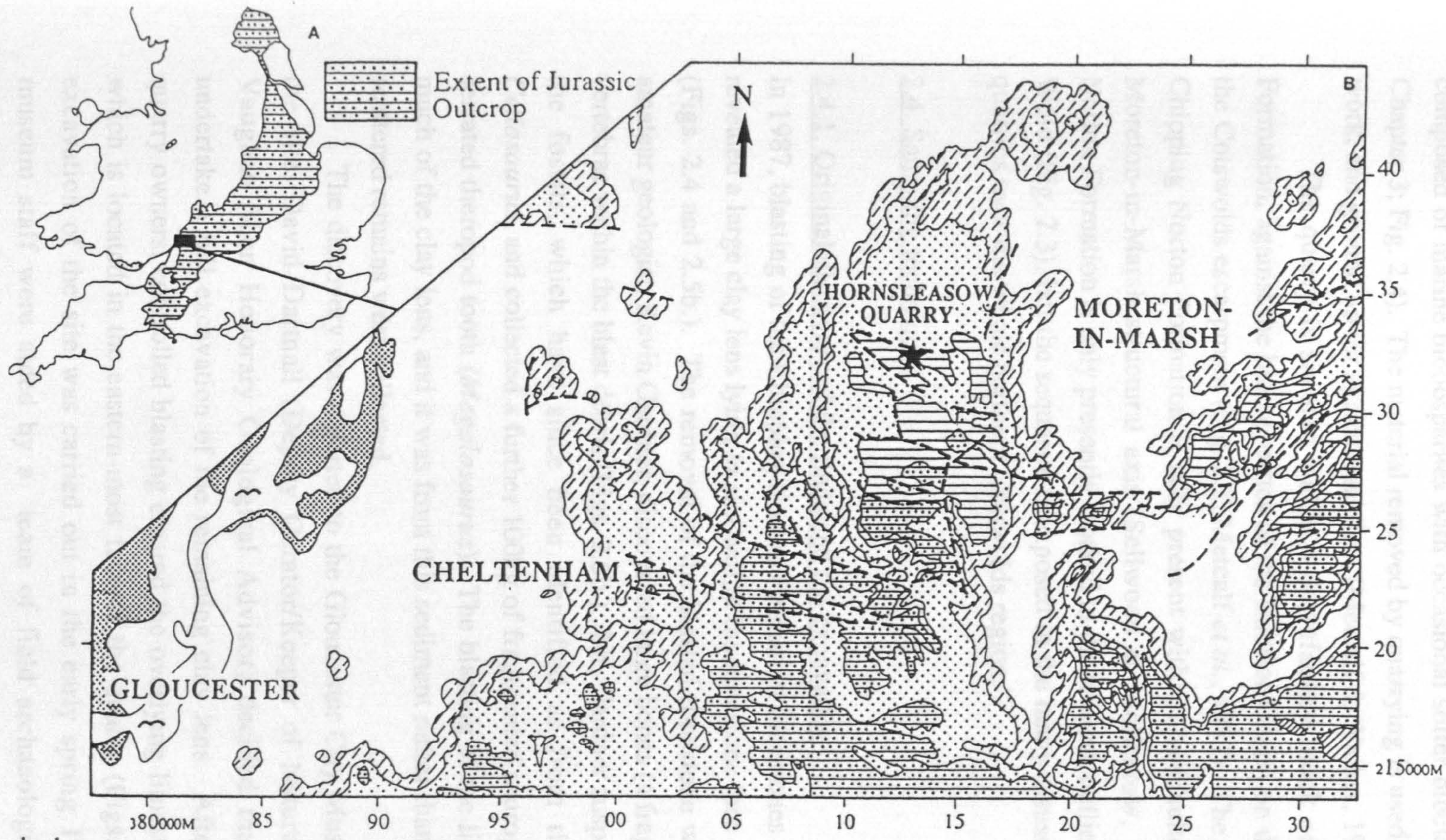


Figure 2.3. Geological map of the northern Cotswolds region, showing location of Hornsleasow Quarry. Inset indicates the extent of Jurassic outcrop in southern Britain. North (N) is indicated.

and is worked occasionally for the Chipping Norton Formation limestones, which are composed of marine biooosparites with occasional softer bioclastic intercalations (see Chapter 3; Fig. 2.4). The material removed by quarrying is used for building and facing work, and ground up for agricultural lime (Metcalf & Walker, 1994).

The quarry is situated within a downfaulted block of the Chipping Norton Formation, against the biooosparites of the Inferior Oolite, on the northeastern edge of the Cotswolds escarpment (Fig. 2.3; Metcalf *et al.*, 1992). The thick succession of the Chipping Norton Formation is only present within downfaulted outliers west of the Moreton-in-Marsh structural axis (Sellwood & McKerrow, 1974). The Chipping Norton Formation is only presently worked in the western outliers around Stow-on-the-Wold (Fig. 2.3), but the sequence is exposed in the many disused and partially infilled quarries over the whole northern Cotswolds region.

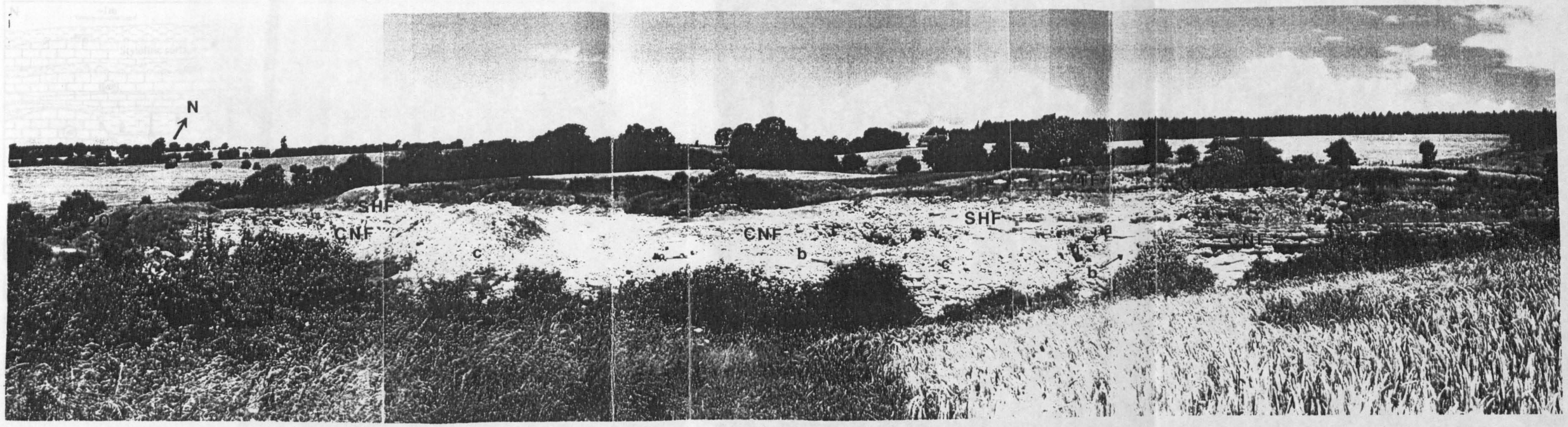
2.4. Sampling techniques

2.4.1. Original discovery of the finds and the excavation

In 1987, blasting of the Chipping Norton Formation limestones by Huntsman's Quarry, revealed a large clay lens lying *in situ* within the limestone face at Hornsleasow Quarry (Figs. 2.4 and 2.5b.). The removal of the blasted limestone was delayed and a local amateur geologist, Kevin Gardner, discovered large bones (a fragmented hip girdle) and vertebrae within the blast debris (Figs. 2.5b.). Mr. Gardner suspected the importance of the fossils, which have since been identified as from the sauropod dinosaur *Cetiosaurus*, and collected a further 100kg of fragmented sauropod bone and one large serrated theropod tooth (*Megalosaurus*). The blasting of the limestone had dislodged much of the clay lens, and it was from this sediment rather than the limestone, that the scattered remains were collected.

The discovery was reported to the Gloucester City Museum, and after visiting the site, David Dartnall (Deputy Curator/Keeper of Natural History) and Roger Vaughan (then Honorary Geological Advisor) decided that the museum would undertake a full excavation of the remaining clay lens. After negotiation with the quarry owners, controlled blasting cleared the overlying limestone from the clay lens which is located in the eastern-most face of the quarry (Figs. 2.4. and 2.6a) and an excavation of the site was carried out in the early spring 1988 (Fig. 2.6b.). The museum staff were aided by a team of field archaeologists, the Crickley Hill Archaeology Trust and a few voluntary workers.

The archaeologists brought with them their own careful field techniques, completely logging every large find and bag of clay removed. The existing face before excavation (Figs. 2.5.) showed a bipartite clay lens about 0.8m thick, and 11m in width. The removal of the overlying limestones of the Chipping Norton Formation was



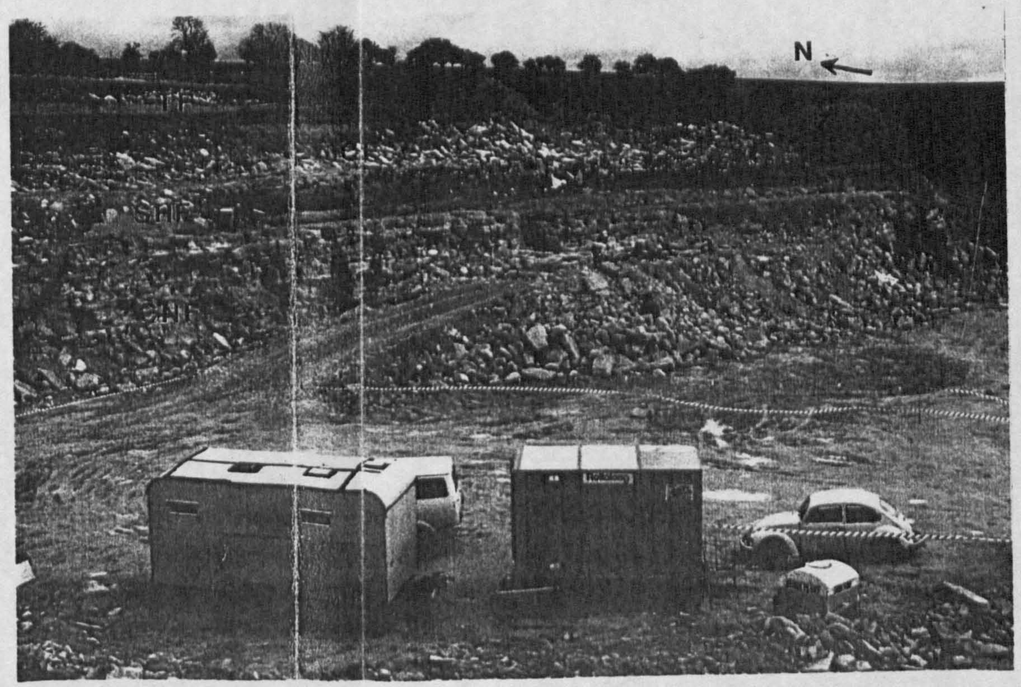


Figure 2.4. Panoramic view (above) looking northwards of Hornsleasow Quarry in 1987, showing the position of the clay lens and excavation (a), waste-heaps of the Hornsleasow clay (b) and of limestone (c). View of eastern part of Hornsleasow Quarry (right) showing the excavation in 1988. In both views north (N) is indicated. Abbreviations for rock formations are: 'CF' Charlbury Formation, 'CNF' Chipping Norton Formation, 'SHF' Sharps Hill Formation and 'TF' Taynton Formation.

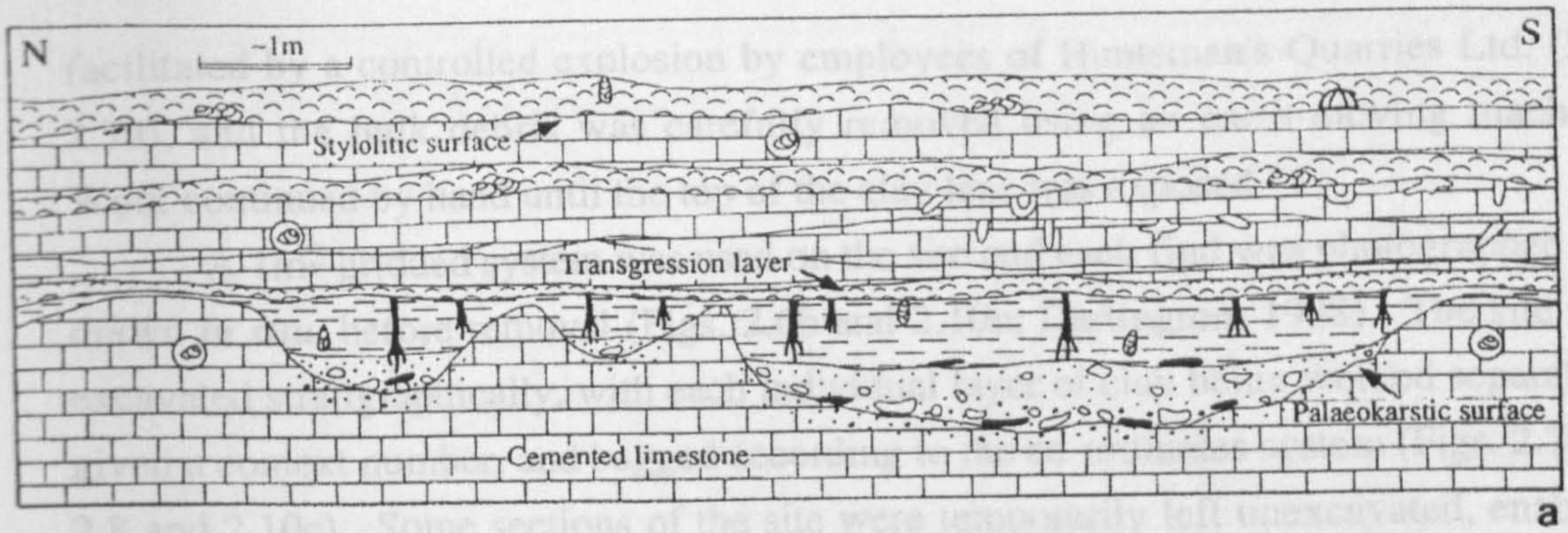


Figure 2.5. An idealised sketch cross-section of the clay lens within the limestone sequence at Hornsleasow Quarry (a), based upon field observations *in situ* and during excavation (after Vaughan, 1989). (b) The Hornsleasow Clay ('HC') unit *in situ*, within the north-east limestone face. Details of the lens include the overlying 'transgressive layer' (1), and the irregular and sharp base (2). (3) indicates the blast rubble from which the original *Cetiosaurus* remains were recovered. Abbreviations as Fig. 2.4.

facilitated by a controlled explosion by employees of Huntsman's Quarries Ltd. (Fig. 2.7a), and the bulk debris was carefully removed using an earth-moving machine. Work continued by hand until the top of the clay lens was exposed.

A $1m^2$ gridded system was used on the site and each find was photographed and drawn *in situ*, before removal (Figs. 2.6b and 2.10b; Darlington, 1988). The site was excavated stratigraphically, with each individual layer of clay being worked separately, given a context number, and bagged according to the co-ordinates system (Figs. 2.7c, d, 2.8 and 2.10c). Some sections of the site were temporarily left unexcavated, enabling detailed drawings to be completed (Fig. 2.9.), showing the exact topography of the lens, and correlation across site of the individual clay contexts.

The lens was seen to continue beneath the face and a further area was blasted to allow a second dig - Site 'B' on the site plan (Figs. 2.6b and 2.9.). This extended the site to a total area of 11.5 x 12m (Fig. 2.6b; although the excavation of 'Area B' destroyed the original excavation site 'Area A'). The upper clay layer, known as context 1021 in 'Area A' and context 1006 in 'Area B' (Fig. 2.9.; Darlington, 1988) was difficult to remove, as when wet, it became sticky and when dry, it set as hard as concrete. Depending on the prevailing weather, it could be removed by trowel, shovel or power chisel. The lower clay layer, described by Darlington (1988) as context 1022 in 'Area A' and context 1007 in 'Area B' was the most fossiliferous of the two clay horizons. All the other sediment contexts are minor variations of the two main clay units (Chapter 3; Fig. 2.9) or the surrounding limestone, and they shall not be described here. An explanation of the context system is described in Darlington (1988).

The large sauropod dinosaur bones (Fig. 2.10c, d) were removed in the standard manner: they were exposed from above, strengthened with 10% paraloid B72 in acetone (a durable and non-yellowing ethyl methacrylate co-polymer consolidant), undercut, covered with a separating layer of tissue paper, cling film or foil, and encased in plaster of paris or by using spray foam (Fig. 2.10d). Many large bones showed considerable shattering. These were photographed and drawn, and each fragment numbered before removal, in order to assist the reconstruction and restoration in the laboratory. Large pieces of wood were either wrapped in cling film and bandaged, or if non-scientific samples, were liberally treated with P.V.A.

Since 1988, the first site has been partly obscured by land slip and rubble from subsequent excavation, but erosion has further exposed the palaeokarstic surface beneath the second site. New finds have been made by prospecting over the excavated area and spoil, including a mammal jaw discovered by Mr. Gardner in late 1990. Huntsman's Quarries Ltd. have allowed an area around the original excavation site to be preserved for further research (Fig. 2.6a.). Further details of the excavation procedures are given by Darlington (1988) and Vaughan (1989), Metcalf *et al.* (1992).

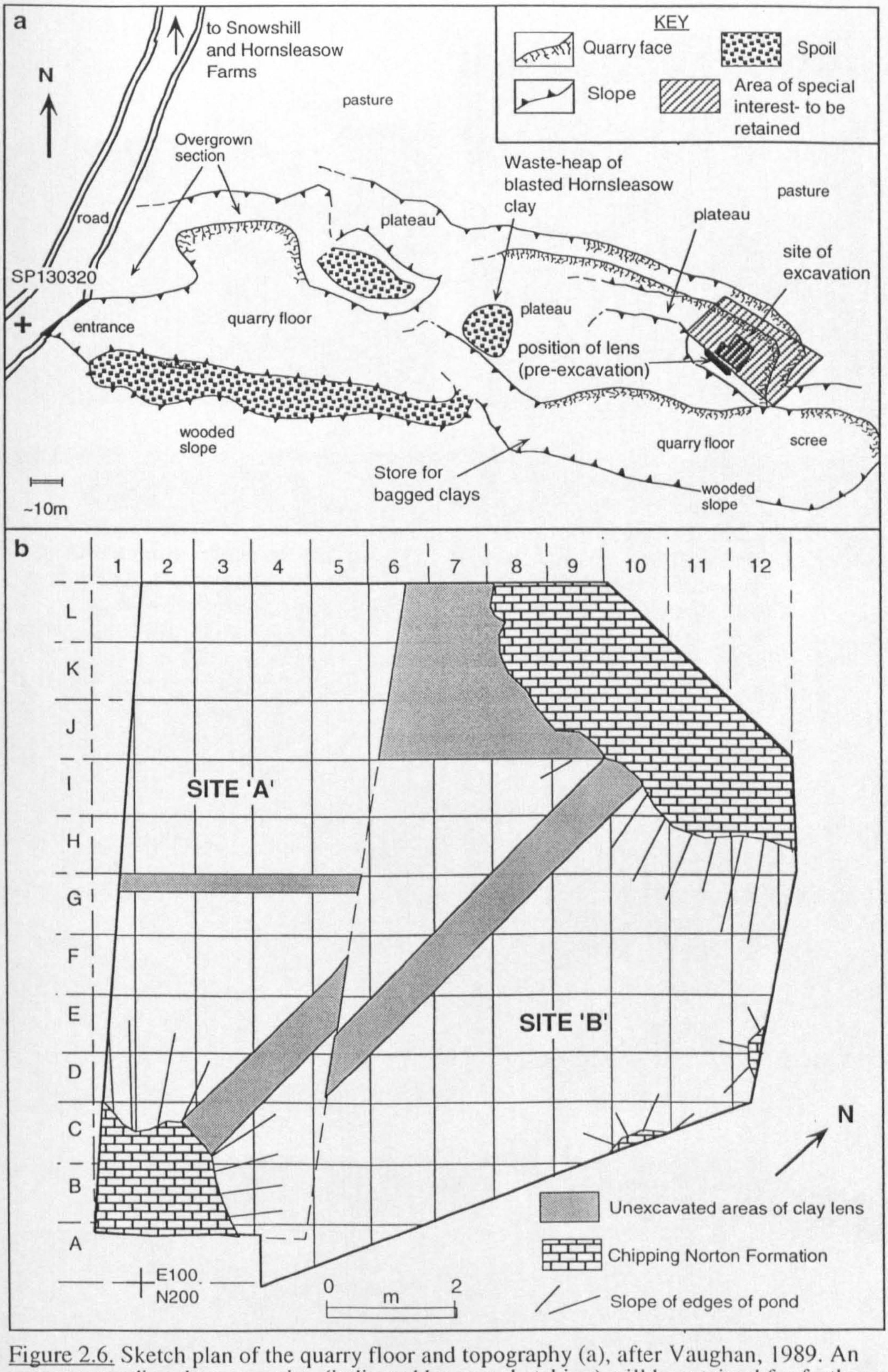
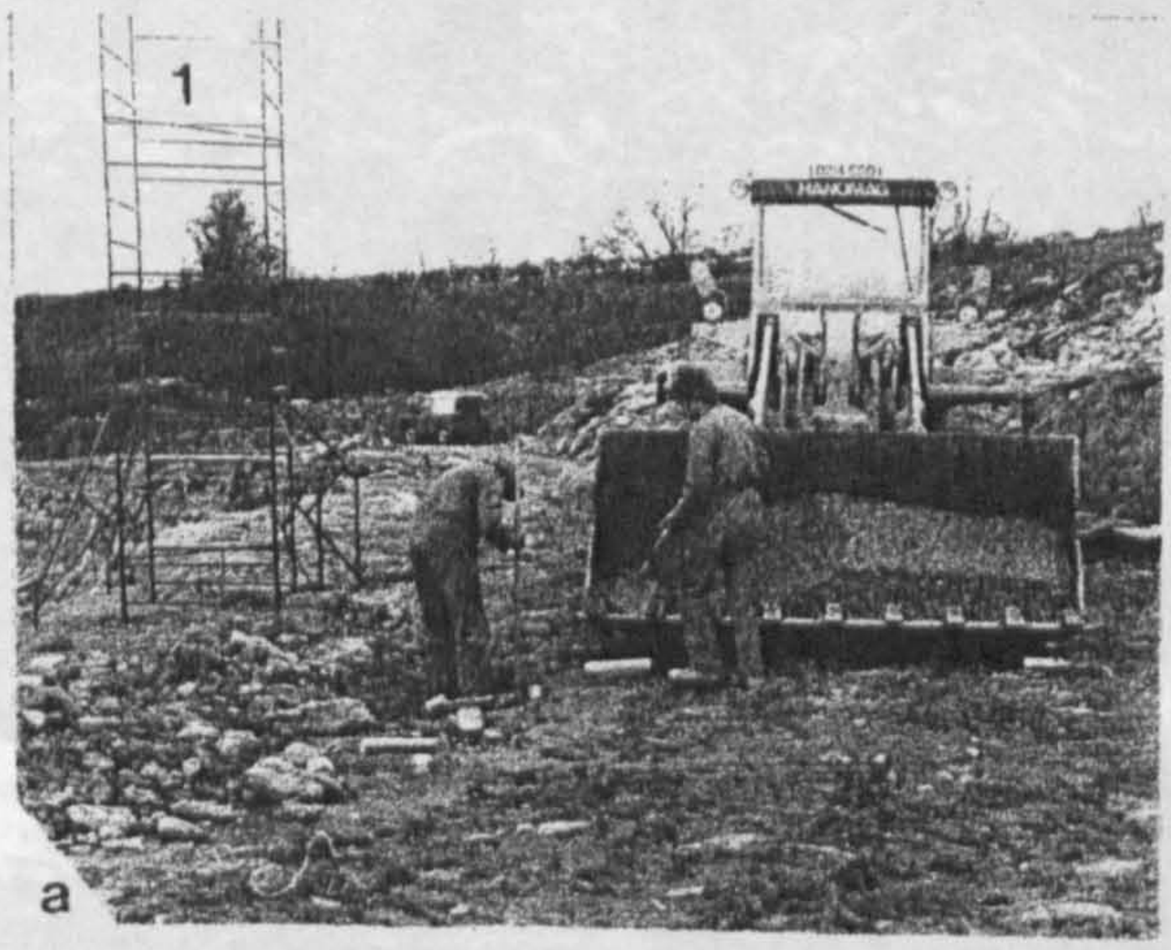


Figure 2.6. Sketch plan of the quarry floor and topography (a), after Vaughan, 1989. An area surrounding the excavation (indicated by cross hatching) will be retained for further investigations. Site plan (b) of the excavation showing the grid system employed on both Site A and B, and the unexcavated sections of the clay lens retained for sedimentological studies. North (N) is indicated.

Figure 2.7. Excavation of the first site, Site 'A' on Fig. 2.5: (a) Clearing overburden (Chipping Norton and Sharps Hill formations) and laying explosive charges, (1) indicates the observation tower used for aerial photographic shots of site ; (b) trial excavation of clay lens to determine contexts (see Fig. 2.9); (c) view from the south of excavation of upper clay contexts (1021), leaving undisturbed vertical clay transects (2) from which details of the internal stratigraphy of the lens were noted for sedimentological analysis (sections A-A' and B-B' on Fig. 2.9). The clay was bagged on site (3) revealing the limestone sides of the clay lens (4); (d) View from observation tower of excavation of lower contexts (1006). Abbreviations as Fig. 2.4.



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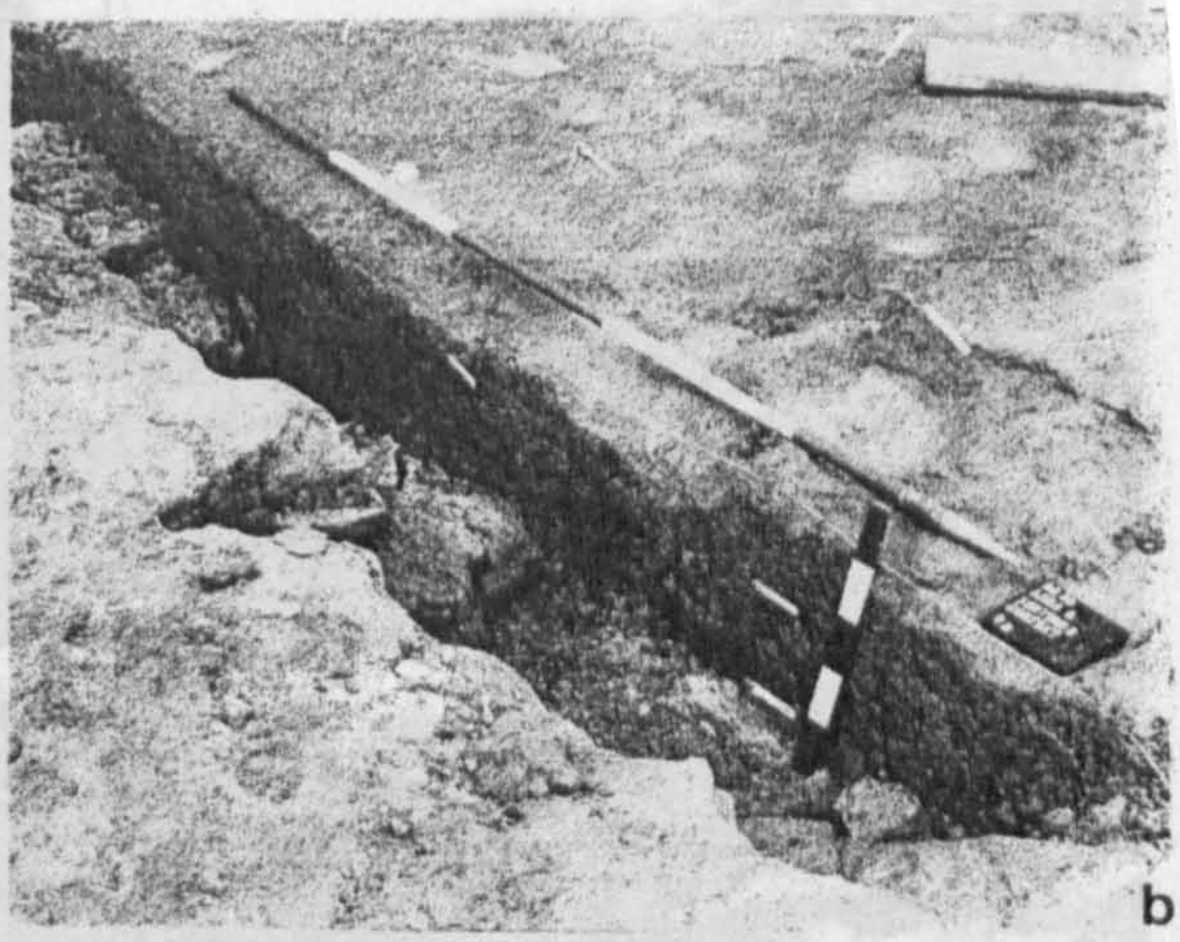

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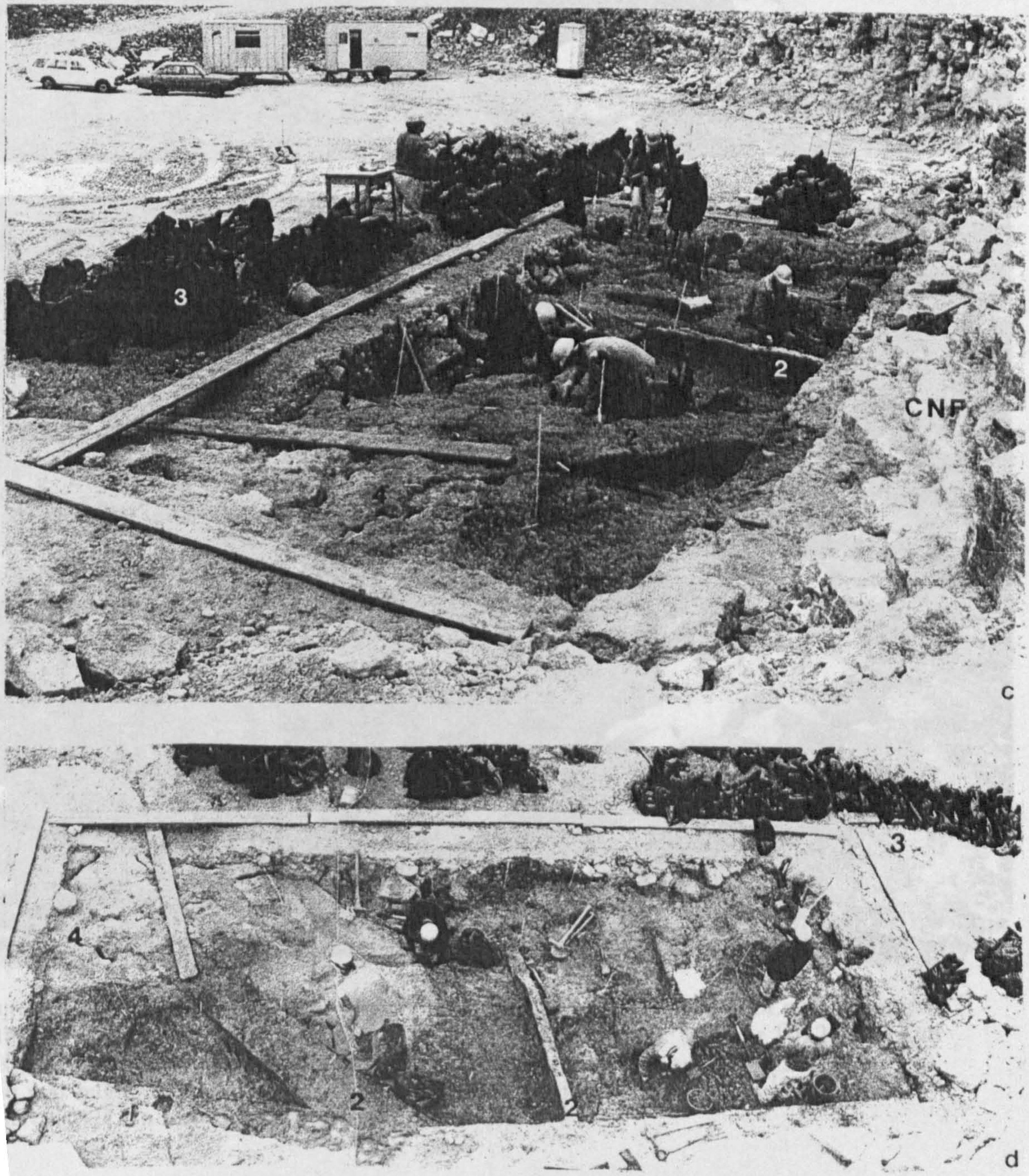

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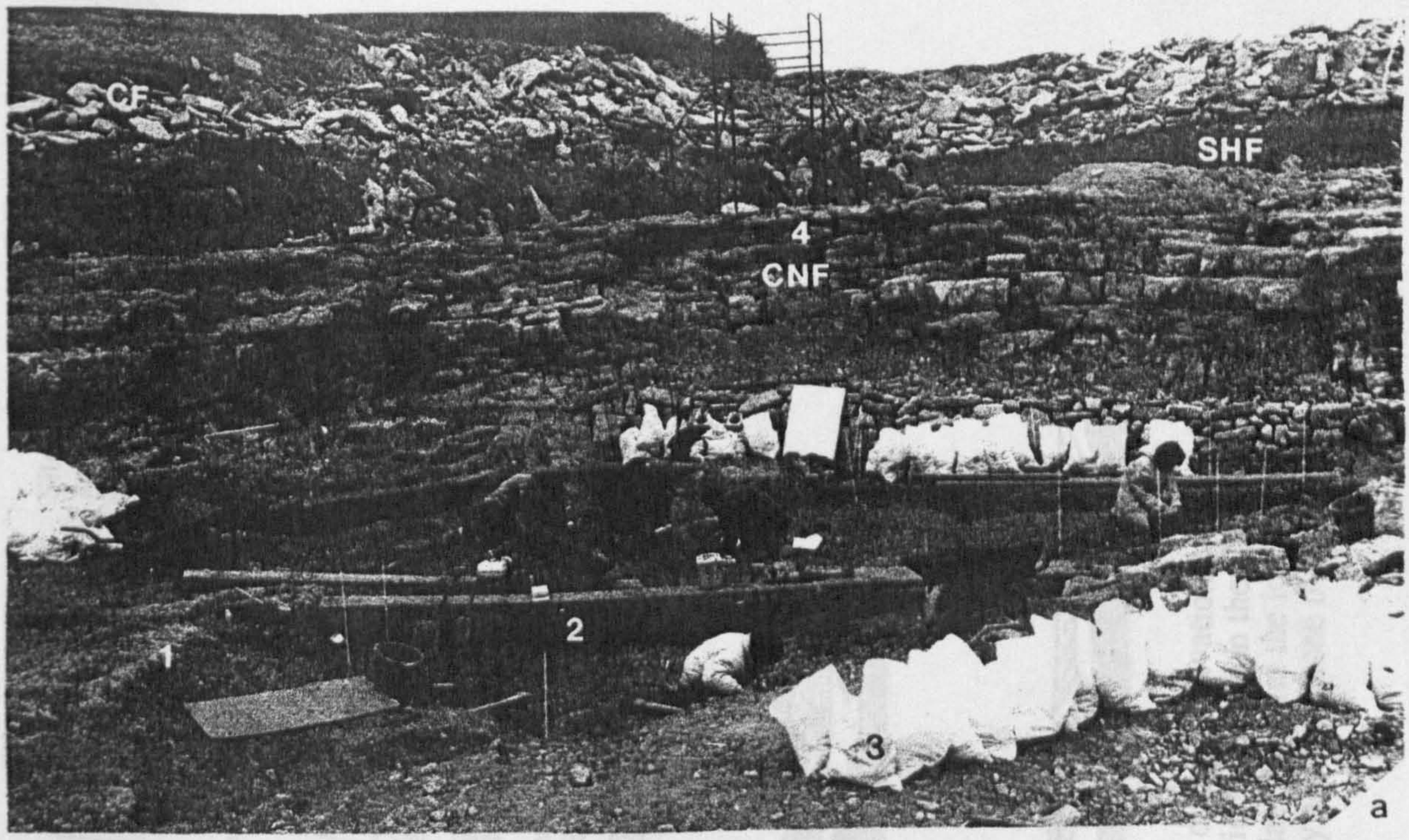


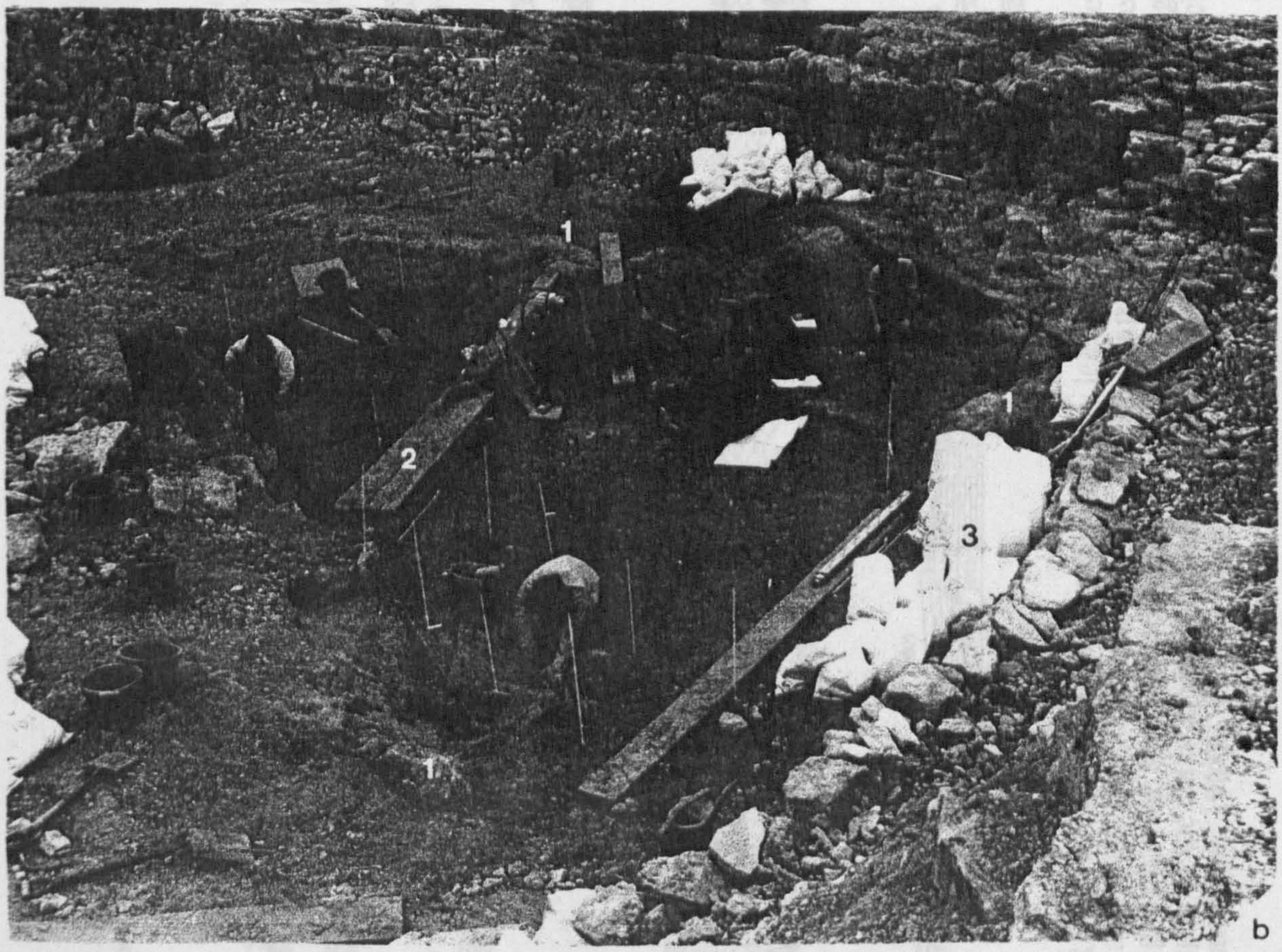


Figure 2.8. Excavation of the second site, 'Site 'B' on Fig. 2.5: (a) view of site looking north-east, showing clearance of one of the lower clay contexts (1007 or 1022) to reveal the sloping limestone edges of the lens (1) and undisturbed section B"-B"'. The clay again was bagged on site (3); (b) view from above of excavation of upper clay units (1006 or 1021) in same area. Abbreviations as Fig. 2.4.

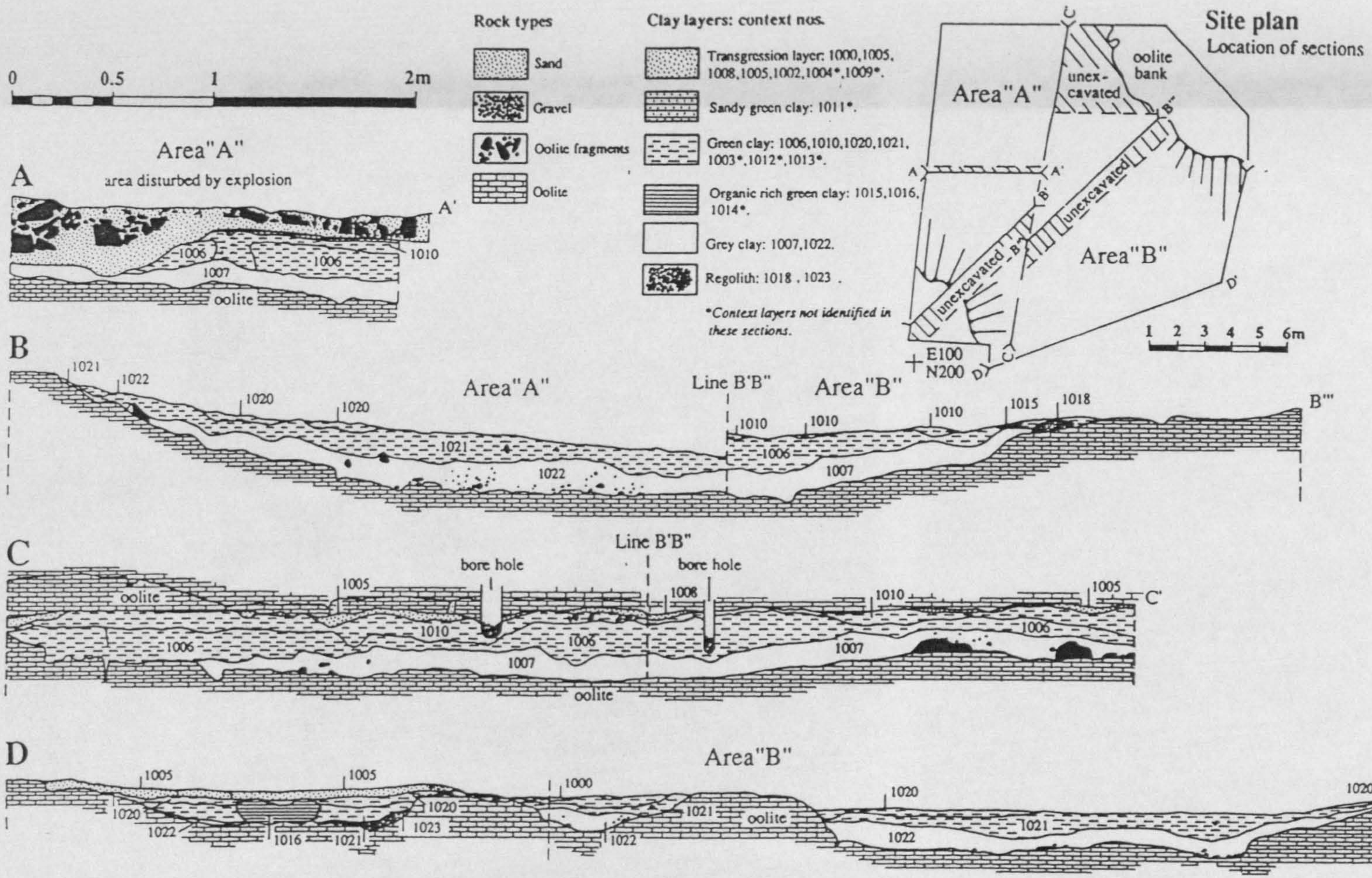
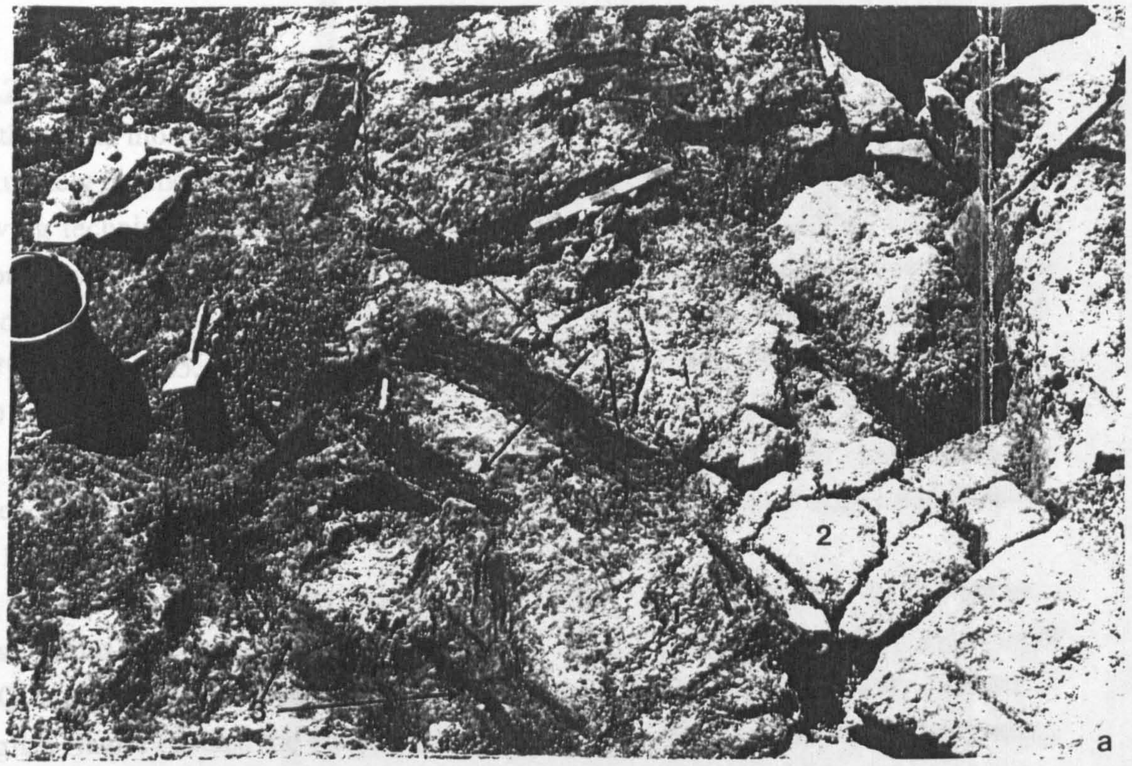


Figure 2.9. Detailed cross-sections across the clay lens based upon data collected during excavation by the archaeologists from undisturbed 'walls' of clay along various lines of intersection (shown on the site plan - inset; Fig. 2.6b). The sections show the general topography of the lens, the irregular base, and the internal stratigraphy of the clays. Context numbers correspond to different varieties of clay and limestone overburden (after Darlington, 1988).



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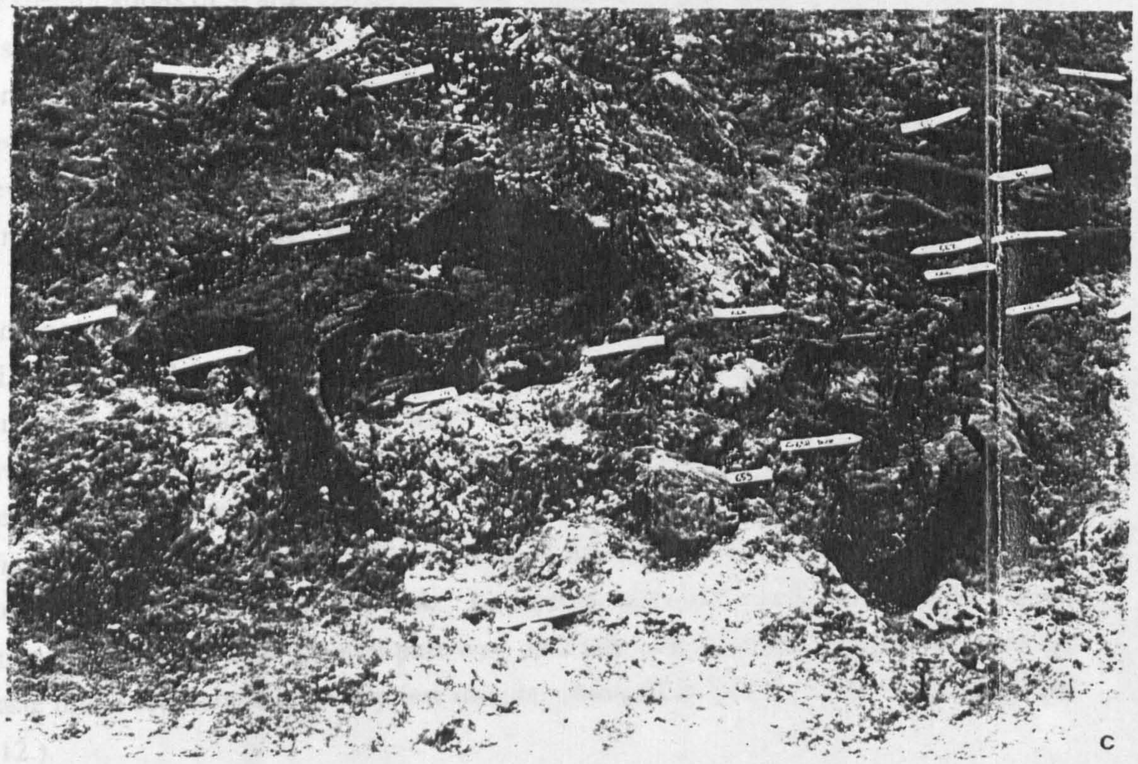

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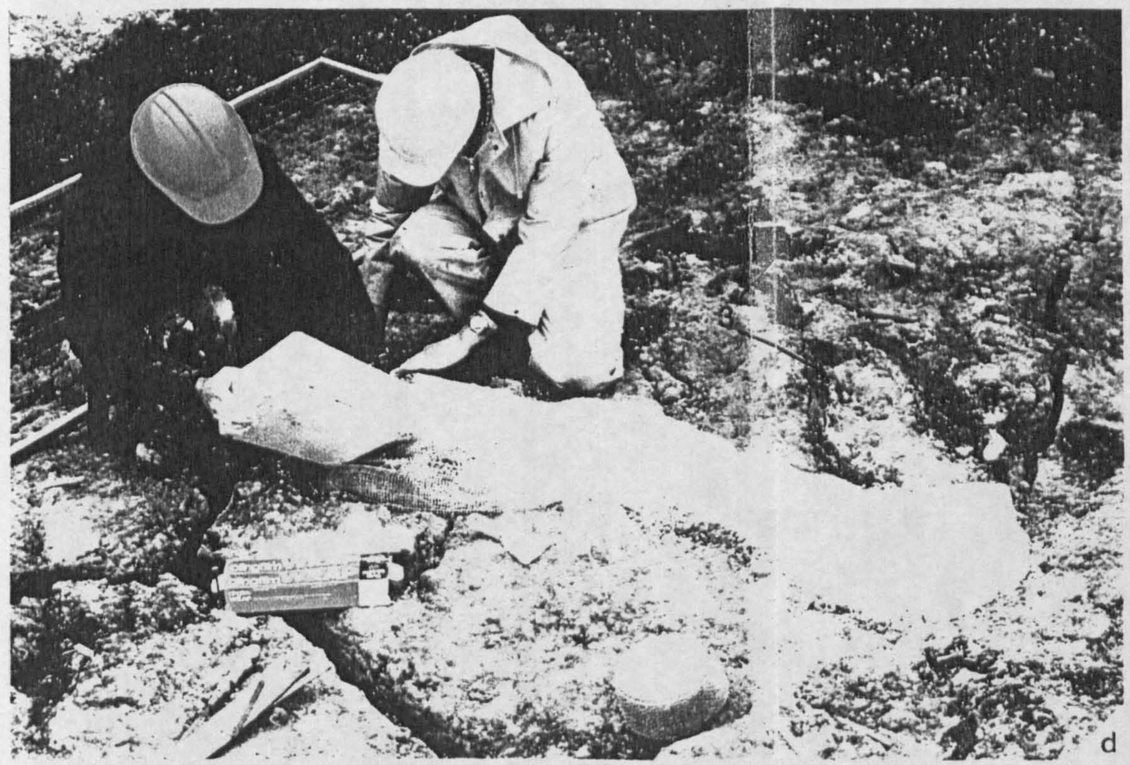


d

Figure 2.10. Excavatory techniques employed on site during the Hornsleasow dinosaur dig: (a) clearing of each clay context individually (1) - this one is the lowest context (1022), to reveal the underlying palaeokarstic surface of the limestone (2) and several large pieces of lignified wood (3) encased within the lower clay; (b) marking position of finds upon the grid system (Fig. 2.6b); (c) labelling and numbering each find on site, before removal. The two labelled bones are part of the *Cetiosaurus* ilium (4) and vertebra (5); (d) applying the outer case of plaster of paris and hessian onto the shattered femur of the *Cetiosaurus*.

2.4.2. The sieving project

In the past, most vertebrate sites were exploited for their large bones only, but fortunately the modern trend is toward searching for all sizes of vertebrate fossils. The sieving technique has long been used to find micro fossils (Kühne, 1946) but its popularity has increased over the past few decades. Recently this type of work has produced many new taxa from the British Triassic and Early Jurassic (Freeman 1976, 1979; Fraser 1985, 1986, 1988; Ensom 1977; Evans *et. al.* 1988; Milner and Evans, 1991). On the advice of Paul Ensom and sieving expert David Ward, a trial sample of the clay was washed for Gloucester Museum. This produced many small bones and crocodile teeth. It was decided that the whole clay lens should be bagged and removed for later washing.

The excavation was completed in 1988, and 23 tonnes of potentially fossiliferous sediment were bagged on the site. Of this, three tonnes were transported to Gloucester City Museum for processing. A grant from the geology section of English Nature enabled the construction of an automated sieving tank at the museum to the specifications of Ward (1981), with some small modifications (Figs. 2.11 and 2.12). A second tank was assembled at Vaughan's home and in October 1990, the project transferred to Bristol University Geology Department under a Natural Environment Research Council grant awarded to Mike Benton. This financed the construction of a sieving machine at Bristol, and the employment of Rachael Walker to process remaining sediments. Since 1990, a further ten tonnes of sediment have been processed.

The sieving tank design consists of a large plastic grit bin with two oscillating, and one rotary, garden sprinklers fitted to the lid into which is set a plastic window (Figs 2.11 and 2.12). Inside the tank are supports for the trays of sediment, which are constructed from plastic bakers' trays with large openings in the sides and bottom covered with fine stainless steel mesh of 350-micron and 430-micron grade. The trays are supported in the tank and remain stationary, while water plays over the sediment (Fig. 2.12.). Clay-grade material is washed through the sieves and settles with the excess water at the base of the tank. At a pre-determined level, the water overflows into an exit siphon tube situated part-way up a side wall, and this flushes out the clay waste into a sediment trap, every few minutes (Metcalf & Walker, 1994; Figs. 2.11 and 2.12.).

To process a tray-load of clay the following procedure is employed. Sediment is tipped into spare trays and left to dry before being sieved, as the completely desiccated clay disaggregates better on washing. The Hornsleasow clays take about seven days to fully dry. The material is then washed for one to four hours, until the clay grade material has been removed. Once dry, the larger pieces of limestone debris and the ferruginous fraction are removed from the residue using a magnifying lens. Smaller

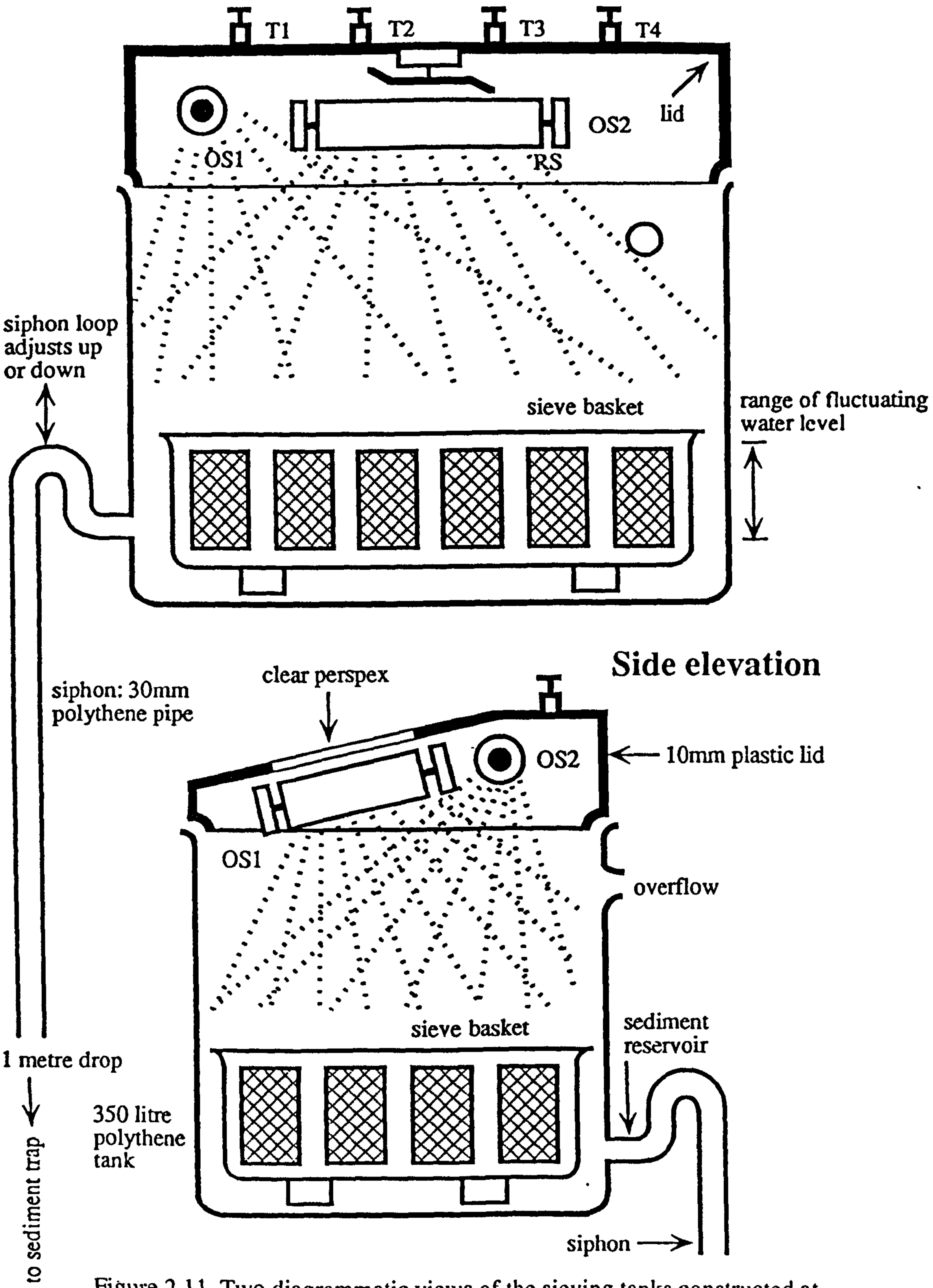


Figure 2.11. Two diagrammatic views of the sieving tanks constructed at Gloucester Museum and Bristol University (after Metcalf & Walker, 1994). Abbreviations for sprinkler system set into the roof of the machine are: 'OS' and 'OS2' oscillatory sprinklers number one and two; 'RS' rotary sprinkler.

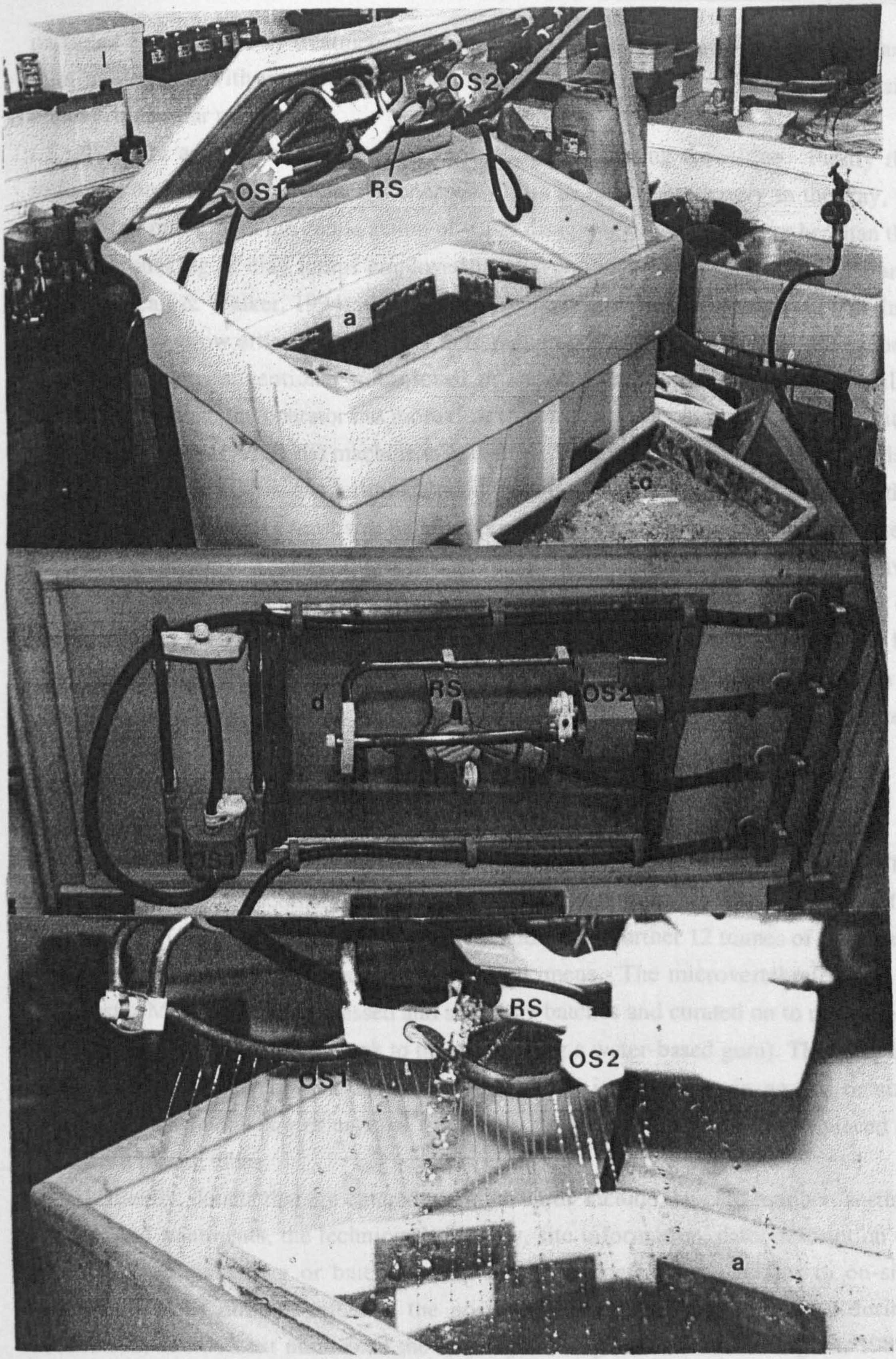


Figure 2.12. Various views of the Gloucester Museum sieving tank, showing trays of sediment within the washing system (a) and undergoing thorough desiccation prior to washing (c), the sediment reservoir pipe (b), and the plastic window set inside the roof of the machine (d). Abbreviations as Fig. 2.11.

particles are removed by treatment in dilute (10%), buffered acetic or formic acid and then neutralised with warm water. The material is then picked over for microfossils under a binocular microscope.

There are two main advantages for using this sieving technique. Firstly the sediment and fragile bone material encased within it, remain stationary in the tray, as the sieving relies upon the gentle action of water disaggregating the clay, rather than the vigorous shaking of clay lumps employed in conventional sieving techniques (Ward, 1981, Metcalf & Walker, 1994). This limits any damage to fragile microfossil remains, and may also allow articulated material to remain in contact (Chapter 10). The second advantage over conventional microfossil processing techniques, is that as a fully automated system, the operator has more time to complete other stages of the extraction and sorting process while the machine is operating. However, there is an increased risk of flooding if the machine is left unattended for any protracted period of time and the clay in the sediment trap must still be laboriously emptied at regular intervals (Walker, pers. comm., 1993). Disadvantages are minimal since the machine is so inexpensive and relatively easy to construct, but the sprinklers and tubing can become rusted or 'furred-up' and parts have to be replaced at regular intervals. The whole process may be slightly slower than conventional methods, but for large loads (i.e. those of over a tonne) and unbiased palaeoecological studies, it is suggested that an automated microfossil sieving tank be constructed.

2.4.3. Curation of specimens

All specimens are retained and curated in the Gloucester City Museum collections. The first three tonnes of sediment yielded over 5000 microvertebrate fossils and 1000 invertebrates (recorded in Vaughan, 1989). In addition, a further 12 tonnes of clay were processed at Bristol, yielding around 19000 specimens. The microvertebrate finds at Gloucester Museum were processed and sorted by batches and curated on to numbered, standard specimen cards (i.e. stuck to the cards using a water-based gum). There are up to 64 specimens per card. Material was not sorted into taxonomic groups or size categories, but interesting or unusual finds were removed from the card and placed in numbered plastic tubes.

Ideally, details that are entered upon the cards include the card number, sorting process and treatments, the technician's identity, site information, date, description of enclosing sediment, bag or batch number and clay context (this relates to on-site information, bag number refers to the position in the gridded system used during excavation, and context number to the clay layer being excavated; Darlington, 1989) and any further comments. These details and identification of the specimen (where possible) have been catalogued on PC FILE database at Gloucester City Museum. Specimens curated in this way include GLRCM G. 50001-66864 inclusive.

The large bone material has been curated separately (Vaughan, 1989). A grant from the Curry Fund (Geologists' Association) enabled conservation and curation of the *Cetiosaurus* remains upon M.O.D.E.S. (Museum Object Documentation Entry System) at Gloucester Museum, in 1990. Details of preparation and conservation of this material appears upon a separate PC FILE, along with field information pertaining to the original position of the bone upon the site plans. Much of the material is fragmentary and therefore not all the specimens have been identified. The catalogue numbers for the *Cetiosaurus* remains and other large skeletal material are GLRCM G. 10000-100171 inclusive.

Since 1990, new finds have been catalogued within a modified Microsoft Excel database, on the Macintosh™ network at Bristol (Metcalf and Walker, 1993). Once again bags of clay were sieved stratigraphically, and batch numbers noted during sorting. However, specimens were curated according to their systematic identity, as well as stratigraphic and site position (i.e. clay context number, batch number and processing details). Identified small finds were placed in glass or plastic 5mm diameter tubes with plastazote foam bung, and given a separate catalogue number. Similar finds were then curated into plastic boxes, which can hold up to 44 individual tubes. For example, one might find up to 44 separately curated, small ornithopod dinosaur teeth within a box.

Larger finds (up to 70mm) were placed in plastic boxes, packed with an inert foam, to prevent breakage. These were also curated systematically, and given individual catalogue numbers. However, as no material was discarded, whatever the preservation, bone or tooth fragments deemed unidentifiable were curated loosely into plastic boxes, according to size. In these boxes there may be up to 100 separate fragments of bone, accordingly a suitable number of catalogue GLRCM "G. numbers" are attributed to the box. This should enable future workers to give a separate catalogue number to any one of these specimens. The material sorted since 1990, has been given the catalogue numbers GLRCM G. 70001-89199 inclusive. No bone material over 70mm has been recovered by the sieving process at Bristol.

The site has been sampled in a painstakingly careful manner, and the excavation and microfossil processing has provided an enormous amount of data for palaeoecological and taphonomic analysis. The site and specimens are still readily accessible for further scientific investigations. The organisation of the Hornsleasow microfossil extraction project has meant that this site has become a model example of a fully documented site on an international scale!

Chapter Three General Geology

3.1. Stratigraphic position of the Hornsleasow vertebrate-bearing horizon within the Chipping Norton Formation

3.1.1. Lithostratigraphy and lateral correlation of the Chipping Norton Formation

Hornsleasow Quarry is a well known field locality for Jurassic stratigraphers. It has been designated an SSSI because of the complete succession of the Sharps Hill Formation (Lower-Middle Bathonian). The position of the Chipping Norton Formation succession at Hornsleasow within the Bathonian stratigraphic column is shown in Fig. 3.1. The sequence at Hornsleasow is described fully in section 3.3. Although the quarry is situated in Gloucestershire the sequence at Hornsleasow is similar to the Oxfordshire Bathonian succession (see Fig. 2.1), and is based upon the work of the British Geological Survey (B.G.S.) in the districts of Moreton-in-Marsh (at the present time the B.G.S. have not completed revision of the memoir for Moreton: Richardson, 1929), and Chipping Norton (Horton *et al.*, 1987).

The 'Chipping Norton Limestone' was first described and named by Hudleston (1878), from a succession at Chipping Norton Workhouse Quarry (SP 319273; Appendix A2) to the southwest of Hornsleasow. He intended the name to apply to the whole of the intervening limestones between the Sharps Hill Formation and the underlying *Clypeus* Grit. The name was formalised to the Chipping Norton Formation or Chipping Norton Limestone Formation by McKerrow & Kennedy (1973), and the type section designated that at the Workhouse Quarry (Sellwood & McKerrow, 1974), although it is now largely overgrown (pers. obs. 1992).

The formation is rather variable in lithology, ranging from a pure white oolitic limestone in the southwest, to a rather sandy deposit eastwards. Four lithofacies have been defined in the Chipping Norton Formation and these were given 'member' status by Sellwood & McKerrow (1974). These lithofacies are :

(1) The Chipping Norton Member. Predominantly bioclastic limestone facies in northwest Oxfordshire and northern Gloucestershire above the *Trigonia signata* Bed;

(2) The Swerford Member. Calcareous arenites at the top of the formation in northeast Oxfordshire.

(3) The Hook Norton Member. Predominantly oolitic freestone in northwest Oxfordshire and northern Gloucestershire below the *T. signata* Bed.

(4) The Roundhill Member. A thin clay seam at the base of the formation in northern Gloucestershire.

Walford (1883) was the first to introduce a bipartite division of the formation, introducing the term 'Hook Norton Limestone' for the part of the 'Chipping Norton Limestone' below the top of the enigmatic *T. signata* Bed (Sellwood & McKerrow, 1974, p.191). The junction bed is a thin, bioclastic oolite with a hardground developed at the top, and is only present in northwest Oxfordshire, and elsewhere the division can be proved only by using palaeontological evidence. The overlying Chipping Norton Member of the Chipping Norton Formation generally comprises fairly soft, thinly-bedded bioclastic limestones and more massive bedded oolites. At Hornsleasow the division is based on lithological characters (section 3.3; Calloman *et al.*, 1993).

In northeast Oxfordshire and in particular at Hook Norton (SP 360320) the formation was split into two beds, with the Hook Norton Member, which is typically a rather even-bedded, massive limestone, with an upper surface of "waterworn, wavy surfaces" known in the earlier literature as the 'Knotty Bed' forming the lower unit (Richardson, 1911, p. 205). The overlying beds in this area are soft, sandy, and fossiliferous oolites and were described by Richardson (1911) as the 'Swerford Beds'. Richardson (1911, 1929), and Arkell (1933), used the term to apply to the whole upper unit: the term 'Chipping Norton Limestone' being used to lump the 'Hook Norton Limestone' and 'Swerford Beds' together. However, the 'Swerford Beds' were formalised as the Swerford Member, but instead of describing the whole of the upper limestone beds *sensu* Richardson, they have been limited to the sands in northeast Oxfordshire (Sellwood & McKerrow, 1974). Torrens (*in* Cope *et al.*, 1980a) provided a summary of British Bathonian stratigraphy (Fig. 3.1). This includes the Swerford Member in the column, as the lowest part of the Chipping Norton Member. The Swerford Member sands are not present at Hornsleasow (Fig. 3.1), but it is possible that the fossiliferous clay lens may be their lateral equivalent (Metcalf *et al.*, 1992; Chapters 5-6).

In northern Gloucestershire, a thin black clay horizon occurs at the base of the Chipping Norton Formation in the railway cutting at Roundhill (SP125221), and was named the 'Roundhill Clay' by Richardson (1929) . It contains a marine fauna and is thought to thicken laterally into the clays of the Lower Fullers Earth Clay to the south (Arkell, 1947b, Sellwood & McKerrow, 1974). It was formalised by Sellwood & McKerrow (1974) as the Roundhill Member, although they admit that it may actually be better considered "as a northern extension of the Lower Fullers Earth [Clay]" (Sellwood & McKerrow, 1974, p.192). The thin clay is not present at Hornsleasow, were the Hook Norton Member limestones overlie the *Clypeus* Grit of the Inferior Oolite (Upper Bajocian; Torrens, 1969a)

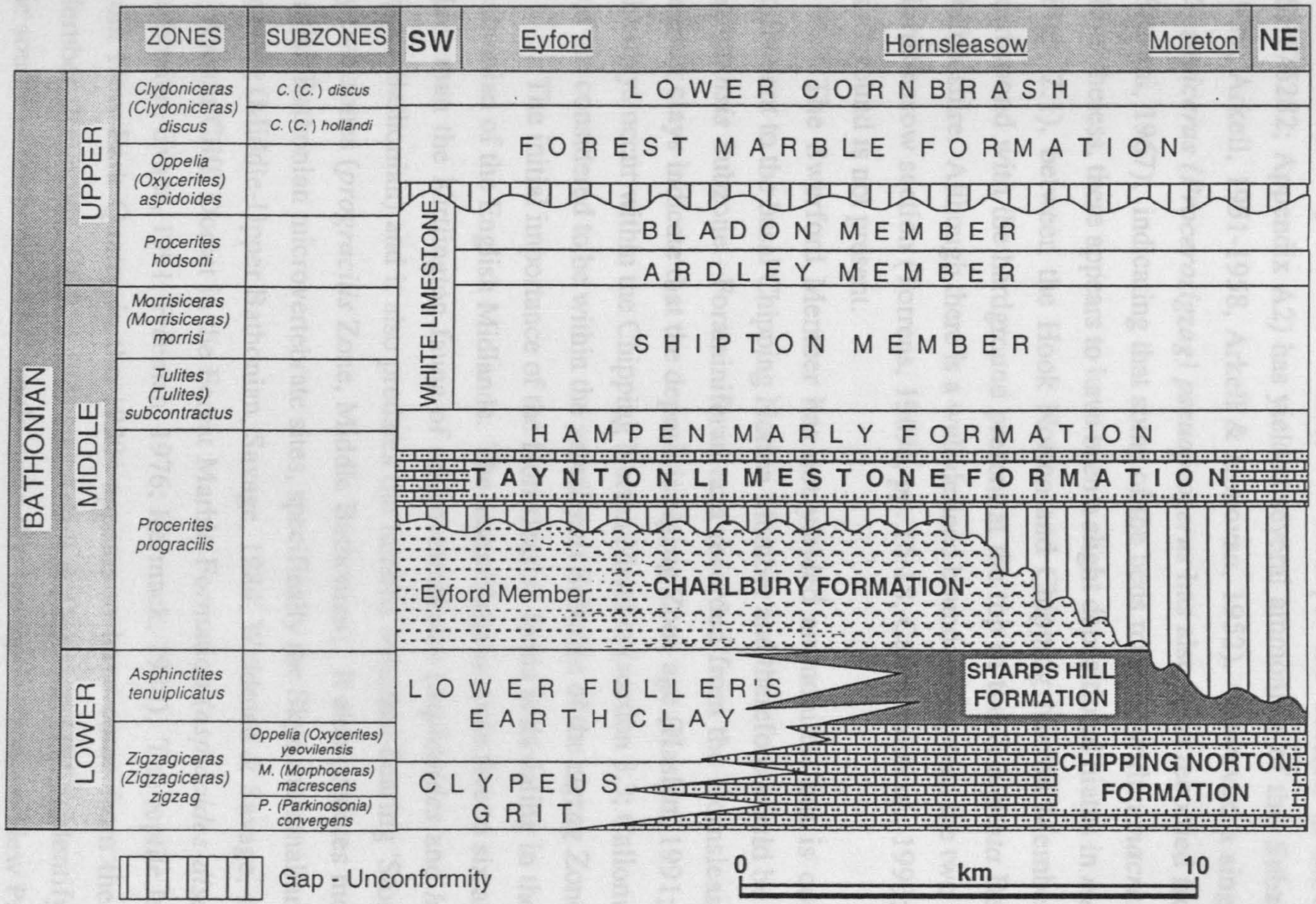


Figure 3.1. Biostratigraphic column and correlation of lithostratigraphic units across the northern Cotswolds for the Bathonian stage. Formations present in the Hornsleasow sections are shaded.

3.1.2 Biostratigraphy

The Chipping Norton Formation oolites are notoriously impoverished in an ammonite fauna (Torrens, 1969b, and *in* Cope *et al.*, 1980a), making precise biostratigraphical correlation of the various lithofacies extremely difficult. However the formation appears to be confined to the *zigzag* Zone of the Lower Bathonian (Cope *et al.*, 1980a; Fig. 3.1). The Chipping Norton Member is assigned to the *yeovilensis* Subzone, based upon the presence of the ammonite *Oppelia* (Torrens, 1969b, p.74). The majority of the ammonites recovered from the Hook Norton Member belong to the *convergens* Subzone (Torrens, *in* Cope *et al.*, 1980a), for example the section at New Park Quarry (SP176282; Appendix A2) has yielded several ammonites of this Subzone (Reynolds, 1939, Arkell, 1951-1958, Arkell & Donovan, 1952). However, a single specimen of *Zigzagiceras (Procerozigzag) pseuoprocerus* has also been recorded from the member (Sturani, 1967), indicating that some of the beds must be within *macrescens* Subzone. Nevertheless, there appears to have been a slight depositional hiatus in *macrescens* times (Figs. 2.1), between the Hook Norton and Chipping Norton Members, which may correspond with the hardground present at the top of the *T. signata* Bed in northwest Oxfordshire. Although there is a well-defined boundary between the two members in the Hornsleasow section (Torrens, 1969a, pp. 16-18, Callomon *et al.*, 1993; section 3.3), a hardground is not present.

The Swerford Member has not yielded ammonites, but is considered to be equivalent to the basal Chipping Norton Member and therefore would be confined to the *yeovilensis* Subzone. Foraminiferan taxa recorded from the Hornsleasow vertebrate-bearing clays indicate that the deposit is *zigzag* Zone age (Haslett, 1991; section 8.2.2). The clays occur within the Chipping Norton Member (section 3.3; Calloman *et al.*, 1993) and are considered to be within the *yeovilensis* Subzone of the *zigzag* Zone.

The initial importance of the Hornsleasow fauna is its dating in the context of the Bathonian of the English Midlands. The assemblage is some five to six ammonite zones older than the Kirtlington fauna of microvertebrates (*aspidoides* and *hodsoni* Zones, Upper Bathonian) and it also predates the famous mammal-bearing 'Stonesfield Slates' by two zones (*progracilis* Zone, Middle Bathonian). It also predates most of the other British Bathonian microvertebrate sites, specifically the Skye mammalian fauna at Loch Scavaig (Middle-Upper Bathonian, Savage, 1984, Waldman & Savage, 1972), and that at Watton Cliff, Dorset in the Forest Marble Formation (*aspidoides-discus* age, Upper Bathonian, Evans, 1991; Freeman, 1976; Kermack, 1988). The reptile fauna recovered from New Park Quarry in the 1930s appears to have come from the Hook Norton Member (Reynolds, 1939, p.193) and Arkell & Donovan (1952) identify these beds as the source of the bones. This would mean that the reptile fauna at New Park Quarry and possibly the bones recovered from Longborough Road Quarry (situated about 500 metres from New Park; SP176292) are older than the Hornsleasow assemblage, although an

earlier reference to vertebrate remains at New Park identified the fossiliferous horizon as occurring within the upper beds in the sequence (i.e. Chipping Norton Member; Richardson, 1929, p.89).

3.2. Palaeogeography

3.2.1. Regional tectonic setting

During the Bathonian stage, southern Britain was covered in the shallow seas of a stable marginal platform basin (Bradshaw *et al.*, 1992). The Great Oolite Group was deposited upon the wide, expansive Cotswold-Weald shelf, which extended north and northeastwards into the shallows of the English midlands (Fig. 3.2a, after Bradshaw *et al.*, 1992). The shelf was bounded to the north, east, and southwest by stable, low-lying land areas (Martin, 1967, Calloman, 1979, Palmer & Jenkyns, 1979, Cope *et al.*, 1992, Figs. 1.5c).

The Cotswolds-Weald shelf is known to have been tectonically active during Lias and early Mid Jurassic times (e.g. Chidlaw, 1987) and fault-block activity in the Early Bathonian may have been responsible for sedimentation patterns (Fig. 3.2b). Several basin and swell structures have been noticed in the succession, and include north-south trending structures with associated sediment thinning, such as the Moreton-in-Marsh swell ('MM' on Fig. 3.2; Hull, 1855), Mendip 'anticline' ('MH' on Fig. 3.2; Arkell, 1933) and Chipping Campden swell (Cox & Trueman, 1920) and northwest-southeast trending 'anticlines' (Buckman, 1901). Arkell (1933) defined these structures as gentle drape folds rather than actual anticlines and termed them "axes of uplift", suggesting that they may have been associated with underlying Palaeozoic structures.

Following geophysical and borehole investigation of the Cotswolds region, Hallam (1958) interpreted the structures as classic basin and swell topography (section 1.3.3; Fig. 3.2b), which are linked with syn-sedimentary movements along extensional faults in the underlying pre-Permian basement. Seismic profiling has largely proved the existence of fault controls for many of the basins and swells in the Cotswolds (Fig. 3.2b) and Whittacker (1985) identified the north-south trending Cotswold-Severn structures as a fault-controlled rift complex influenced by the underlying 'Malvernoid' trend (Fig. 3.2a). The downthrow upon these normal faults are generally toward the west (Williams & Whittacker, 1974) and they are thought to have been influenced by the evolving North Atlantic Rift system (section 1.3.3).

The faults in this region actually displace the Triassic (Audley-Charles, 1970) and Lower Lias (Kellaway & Welsh, 1948), but disappear under the later Jurassic strata, although the trends continue as axes of swells and influenced late Lias (Chidlaw, 1987) and Inferior Oolite sedimentation (Mudge, 1978, Baker, 1981). Horsts such as the Mendip and Moreton-in-Marsh swell may have been land areas at certain times in the Lias (Chidlaw, 1987). The Hornsleasow palaeokarst lies upon the flanks of the Moreton

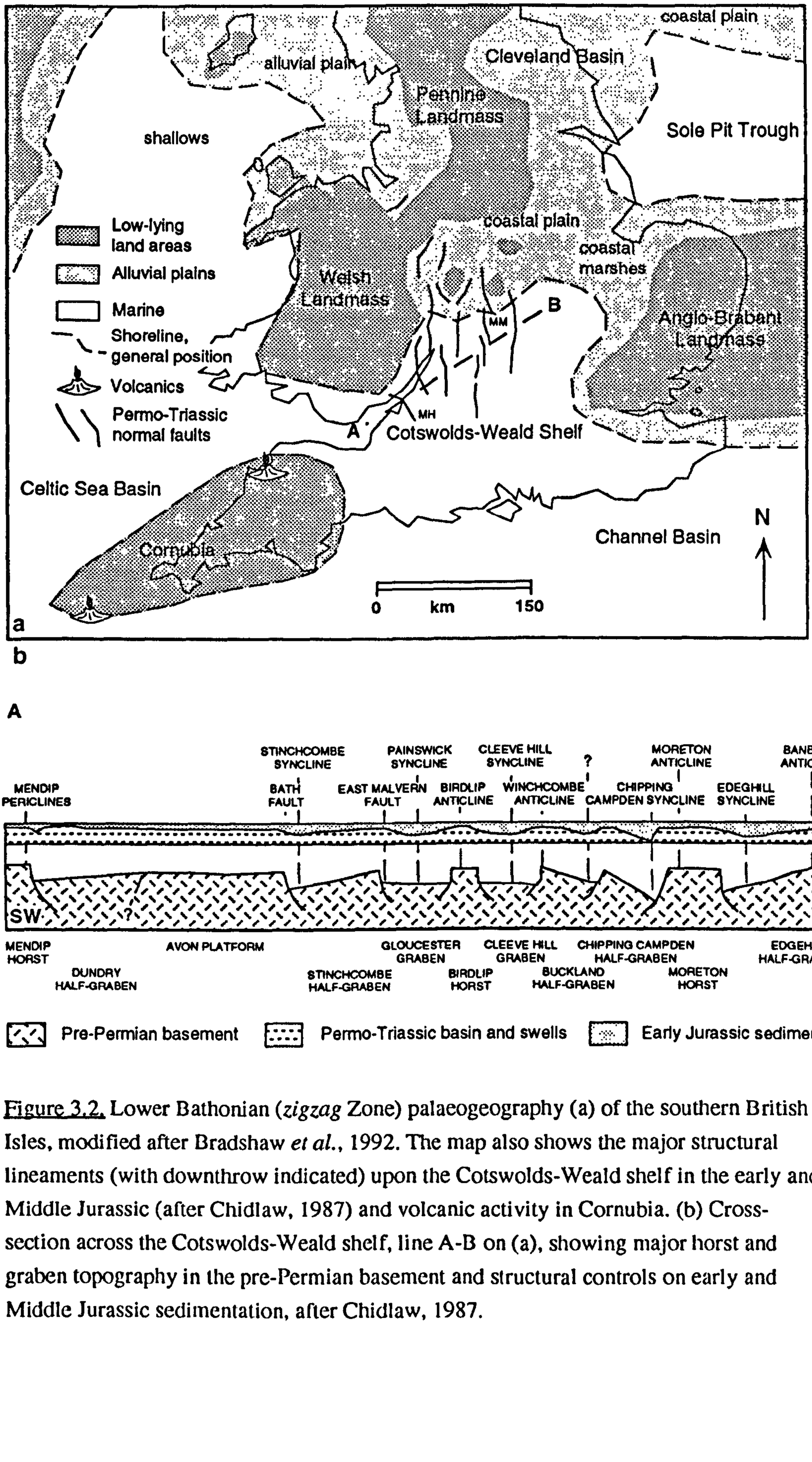


Figure 3.2. Lower Bathonian (*zigzag* Zone) palaeogeography (a) of the southern British Isles, modified after Bradshaw *et al.*, 1992. The map also shows the major structural lineaments (with downthrow indicated) upon the Cotswolds-Weald shelf in the early and Middle Jurassic (after Chidlaw, 1987) and volcanic activity in Cornubia. (b) Cross-section across the Cotswolds-Weald shelf, line A-B on (a), showing major horst and graben topography in the pre-Permian basement and structural controls on early and Middle Jurassic sedimentation, after Chidlaw, 1987.

swell, over which the Chipping Norton Formation thins substantially, suggesting that this may have become an island in early Bathonian times (Chapter 6). By the Lower Cornbrash (*discus* Zone, Upper Bathonian), the tectonic control for sedimentation in the region had ceased (Chidlaw & Campbell, 1988).

3.2.2. Regional patterns in sedimentation

The Bathonian stage in the British Isles is a relatively regressive phase in the primarily marine strata of the Jurassic. Sea-level curves suggest that there was a stillstand in the early Bathonian (Section 1.2.6), with a steady but slow transgressive phase toward the end of the stage, culminating in the fully marine faunas of the Lower Cornbrash (*discus* Zone, Upper Bathonian) and overlying Callovian sequence.

The Chipping Norton Formation oolites attain their maximum thickness around Chipping Norton, northwest Oxfordshire, where they form an extensive plateau and thin steadily toward the northeast and east, before dying out completely in east Oxfordshire. Toward the west and southwest around Cirencester, the formation passes into the quiet deep marine clays of the Lower Fullers Earth (Horton *et al.*, 1987). The beds thin considerably over the Moreton-in-Marsh swell (Sellwood & McKerrow, 1974). In the extreme north and also close to its eastern limit, the Chipping Norton Formation overlaps the Clypeus Grit, to rest upon horizons as low as the Upper Lias (Horton *et al.*, 1987).

The biooosparites of the northern Cotswolds thins steadily and becomes more sandy to the northeast and east of the Chipping Norton district (British Geological Survey, Sheet 218) around Swerford and Great Tew (Swerford Member). These sandy limestones contain up to 30% terrigenous siliciclastic material. They may eventually grade into the marginal marine facies of the Rutland Formation in east Oxfordshire and Northamptonshire fringing the London-Brabant massif (Horton, 1977, Fenton *et al.*, 1994), but as they have limited lateral extent and lack good sedimentary structures, this is debatable (see Chapter 6). The Chipping Norton Formation limestones accumulated as landward prograding oolite shoals in a high-energy, shallow water environment (Sellwood & McKerrow, 1974). The shoreline appears to be largely in the north (?Pennine High) and northeast, although terrigenous material may have also been transported from the Moreton swell.

Palaeoenvironmental reconstructions of many of the formations within the Great Oolite Group of the Cotswolds indicate Florida-type restricted marginal marine and lagoonal conditions, with fresh-water and terrestrial influence at times (e.g. Palmer, 1979, Sellwood & McKerrow, 1974). Periodic exposure of the carbonate shelf is also indicated by the presence of supratidal and karstic features in some of the Great Oolite limestones fringing the London-Brabant landmass (Palmer, 1972, 1979). However, the Hornsleasow deposit is some 50-60 km away from the island in the east and the shoreline of the northern landmass (Pennine) is highly conjectural for the Lower Bathonian (Bradshaw *et al.*, 1992). It is more probable that the terrestrial conditions at

Hornsleasow were fairly localised in an area fringing the Moreton-in-Marsh swell. Possible alluvial deposits contemporaneous to the Hornsleasow clays occur in northeast Oxfordshire in an area known as the 'Midlands shallows' (Chapters 3-6; Hallam, 1992) and some connection with a more substantial land area is indicated by the megafauna discovered at the site (Chapter 9).

3.3. The section at Hornsleasow Quarry

3.3.1. Previous work

The earliest record of the section was made by L. Richardson for the British Geological Survey in 1929, who mentions three old workings situated to the south of the present faces (Richardson, 1929, p.89). In describing the 'Chipping Norton Limestone' (cf. Chipping Norton Formation: Cope *et al.*, 1980), he mentions that it is "conspicuously current-bedded...[and] contains locally pebbles of oolite and black specks". Arkell (1947, p. 37) has described the 'black specks' as "due to ground-up lignite". Richardson (1929, p. 89) also mentions the presence of the *Clypeus* Grit (an oolitic limestone of the Cotswolds, famed for its echinoid fauna, including *Clypeus plotii*, after which it was named) below the Chipping Norton Formation.

Channon (1950) was the next to describe the succession at the then 'Snowshill Hill Quarry'. The workings had been greatly extended since Richardson recorded his sections, and Channon observed that the full succession of the 'Sharps Hill Beds' (c.f. Sharps Hill Member: Cope *et al.*, 1980a) was now exposed. He tentatively correlated his section with Arkell's (1947, p. 63) section from Sharps Hill Quarry (SP 338358) in Oxfordshire. Channon (1950, p. 249) also described the strata directly above and below the 'Sharps Hill Beds': he recognised above a "hard, brown and cream coloured, fissile oolitic limestone", which he believed to be the lateral equivalent of the 'Stonesfield Slate' of north Oxfordshire, and below them he records the 'Chipping Norton' and 'Hook Norton Limestones' (Chipping Norton and Hook Norton Members: Cope *et al.*, 1980a).

Channon separated the two beds at Hornsleasow, but did not retain the term 'Swerford Beds' for the upper limestones, instead he applied the term 'Chipping Norton Limestone' for the uppermost beds. The division was on the grounds that at the junction there was a "large lenticle of tough black clay, barren and extending for about a third of the length of the face, dying out at each end and with a maximum thickness of one foot" (Channon, 1950, p. 249). Quite clearly this previous clay lens is at the same level stratigraphically as the clay bed currently exposed at Hornsleasow.

The section at Hornsleasow was most fully described by Torrens (1968, p. 253 1969) after quarrying had extended the sequence. Although he accepted Channon's division of the Chipping Norton and Hook Norton Members, Torrens could not establish the junction because of the lack of palaeontological evidence. Torrens described the work of identification of the separate lithological units in the quarry as "not easy"; nevertheless,

working with M.J. Barker, he produced a detailed section with five separate lithologies (summarised below after Torrens 1969, pp. 16-18; for full section see Appendix B1):-

5. TAYNTON STONE: 2.5-3.0m [Taynton Limestone; Cope *et al.*, 1980].

4. SHARPS HILL BEDS: 5.48m [Sharps Hill Formation; Cope *et al.*, 1980]

3. CHIPPING NORTON LIMESTONE

Massive, shelly, cross-bedded, oolitic limestone with intercalated softer, orange coloured, oolitic marls. Basal bed is more fossiliferous than the rest, with abundant *Liostrea [Praeexogyra* sp. and *Clypeus* sp., with echinoid spines, *Trigonia* sp., *Lima [Plagiostoma] cardiformis* J. Sow., and *Strophodus* teeth. Finely preserved trace fossils (*Rhizocorallium)* also preserved: 1.5-2m.

2. HOOK NORTON LIMESTONE

Massive, evenly-bedded, hard, oolitic limestone with softer intercalations, cross-bedded in places. Very unfossiliferous and only oyster shell debris and *Lima [Plagiostoma]* cf. *cardiformis* J. Sow: seen c. 10m.

Gap - strata obscured - junction not visible.

1. CLYPEUS GRIT: seen 5m

The upper beds of the Sharps Hill Formation called by Channon (1950) the 'Stonesfield Slates' (his Bed 7) and described in Torrens' section as flaggy arenaceous limestones (his Beds 4?g and 4?i), with intercalated soft oolitic marls (Beds 4?h and 4?j) (Appendix B1) are now equated with the newly defined Charlbury Formation (Boneham & Wyatt, 1993) by this author (see below). It is also probable that the very lowest part of the Taynton Formation, described by Torrens as extremely fossiliferous, rippled marked and as a flaggy buff, oolitic limestone (Bed 5, Appendix B1), may be the lateral equivalent of the *Rhynchonella* Bed-Sevenhampton Marl which is persistently seen at the boundary between the biomicrites and flaggy limestones of the Charlbury Formation and the overlying clean biooosparites of the Taynton Formation (pers. obs.). However, as much of this part of the section at Hornsleasow has now degraded and is overgrown, the junction between the two formations is unclear (pers. obs.).

3.3.2. This work

Following the discovery of the fossiliferous clay lens in the section at Hornsleasow, much work has been completed on the detailed stratigraphy of the sequence in the quarry, including field logs (Fig. 3.3) and descriptions by R.F. Vaughan (1988-1989, e.g. Vaughan, 1989) and myself and others (Fig. 3.4; e.g. Callomon *et al.*, 1993). I have not followed Channon (1950) in placing the present clay lens at the junction between the Hook Norton and Chipping Norton Members, although the temptation would be to correlate the palaeokarstic surface seen below the clays at Hornsleasow (Fig 3.3) with the non-sequence boundary placed above the Hook Norton Member within the Oxford sequence by Cope *et al.*, (1980) (Figs. 2.1 and 3.1). Recent work by N.J. Chidlaw (pers. comm., 1993) corroborates this view in terms of a lithostratigraphic subdivision of

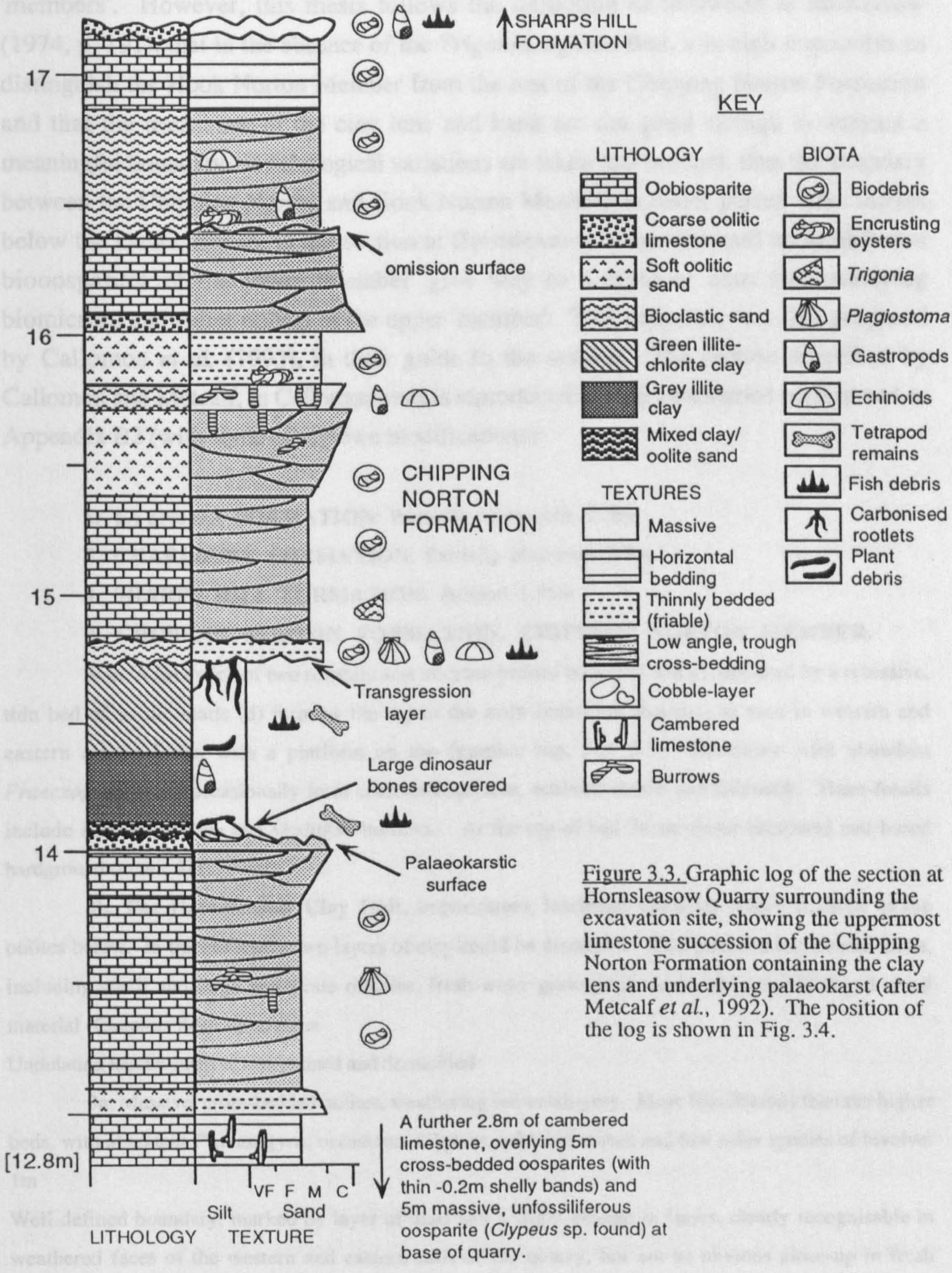


Figure 3.3. Graphic log of the section at Hornsleasow Quarry surrounding the excavation site, showing the uppermost limestone succession of the Chipping Norton Formation containing the clay lens and underlying palaeokarst (after Metcalf *et al.*, 1992). The position of the log is shown in Fig. 3.4.

the two members. Chidlaw argues that the underlying limestones are lithologically distinct from those overlying the clay lens at Hornsleasow and therefore are justifiable 'members'. However, this thesis follows the definition of Sellwood & McKerrow (1974, p. 191), that in the absence of the *Trigonia signata* Bed, it is nigh impossible to distinguish the Hook Norton Member from the rest of the Chipping Norton Formation and that the transience of the clay lens and karst are not good enough to warrant a meaningful boundary. If lithological variations are taken into account, then the boundary between the Chipping Norton and Hook Norton Member, is better placed three metres below the karstic surface in the section at Hornsleasow, where the hard unfossiliferous biooosparites of the lower 'member' give way to a layer of marl and overlying biomicrites and softer oolites of the upper 'member'. This definition was first proposed by Callomon *et al.* (1993), in their guide to the section. The section described by Callomon and Metcalf, in Callomon *et al* is reproduced here in a shortened version and as Appendix B2 (with some of my own modifications):

6. TAYNTON FORMATION: Partially overgrown, 2-3m.

5. CHARLBURY FORMATION: Partially obscured, 3.7m.

4. SHARPS HILL FORMATION: Around 1.75m.

3. CHIPPING NORTON FORMATION. CHIPPING NORTON MEMBER.

3c-e. A sandwich of two resistant sets of cross-bedded oolites (c and e) separated by a recessive, thin bed of oolitic marls (d) forming the top of the main limestone sequence as seen in western and eastern quarry faces, with a platform on top (graphic log, Fig. 3.3). Bioclastic with abundant *Praeexogyra* which occasionally form encrusted surfaces, echinoid debris and fish teeth. Trace-fossils include *Rhizocorallium* and *Skolithos* burrows. At the top of bed 3e an oyster-encrusted and bored hardground occurs: 2m

3b. **The Hornsleasow Clay Unit.** Impersistent, lenticular clays set within hollows in the oolites below. In thicker lenses two layers of clay could be discerned. Well-preserved terrestrial fauna, including small and large vertebrate remains, fresh-water gastropods, ostracods, coprolites and wood material (Chapters 8-9): up to 0.8m

Undulating karstic surface, iron-stained and decalcified

3a. Massive, cross-bedded oolites, weathering brownish-grey. More fossiliferous than the higher beds, with abundant *Praeexogyra*, occasional *Clypeus*, echinoid spines and few other species of bivalve: 1m

Well-defined boundary, marked by layer of marl and a slight change in facies, clearly recognisable in weathered faces of the western and eastern ends of the quarry, but not so obvious close-up in fresh sections. The beds below weather to a more golden-yellow colour.

2. CHIPPING NORTON FORMATION. HOOK NORTON MEMBER. Massive, hard oolitic limestones in many well-bedded courses, which individually are often more or less cross-bedded. very sparsely fossiliferous, oyster debris being the main discernible macrofossil: total thickness now visible 7m (Formerly 10.8m seen, N.J. Chidlaw, pers. comm.).

1. CLYPEUS GRIT. Partially overgrown, formerly seen 5m.

In this description the Chipping Norton Member is defined as more or less massive biooosparites, separated by sharp surfaces (such as surfaces of non-deposition and hardgrounds) or softer intercalations of oolitic marls. Most of the harder limestone units are planar cross-bedded and several show fine vertical burrows levelled at the top by the foreset surfaces of subsequent cross-beds. The section was described around the excavation by Metcalf *et al.* (1992) (Figs. 3.3 and 3.4), but details change rapidly along the face (Fig.3.5a). A five-fold subdivision appears to be persistently discernible and was incorporated into the definition to aid interpretation of the section (Callomon *et al.*, 1993).

3.3.3 Graphic log around fossiliferous horizon

Soon after the discovery, Vaughan (1989) produced a detailed log of the section immediately surrounding the clay lens (Fig. 3.4), which was extended and improved upon by this author (Fig. 3.3, after Metcalf *et al.*, 1992). The divisions used by Torrens (1968), have been employed in the production of the graphic log (Fig. 3.3), showing the clay unit to lie approximately 2.5m below the base of the Sharps Hill Formation, within the Chipping Norton Formation. A transgressive cross-bedded, sandy layer is also seen directly overlying the clay lens (Fig. 3.3) and where the clay has pinched out, the sand unit overlies the palaeokarstic surface (Fig. 3.5a, b). This is a carbonate-rich marine sand, rather than the land-derived silica sands of the Swerford Member, and the two do not appear to be laterally equivalent (section 5.14). However, the clays contain a fairly high percentage of fine-fraction silica sand and may be the lateral equivalents of the Swerford Member sands (section 6.1).

3.4. The Chipping Norton Formation limestones

The lithology of the Chipping Norton Formation is extremely laterally variable over the extent of outcrop (section 3.1.1). It ranges from a pure oosparite in north Gloucestershire and northwest Oxfordshire to calcareous sands in northeast Oxfordshire. To the south and west of this area, the formation interdigitates and eventually gives way to the deep marine argillaceous deposits of the Fullers Earth Clay. The facies variations and depositional environment for the Chipping Norton Formation over the northern Cotswolds is considered in more detail in Chapter 6.

In the type area around the Moreton axis the formation is, however, typified by sandy oolitic limestones with occasional shelly bands (Skerl, 1925). The oolitic limestones range from fine to medium grained grainstones (Dunham, 1962; Table C1, Appendix C), whilst the shell-fragmental bands tend to be grainstones or rudstones (Dunham, 1962). Compositionally, the limestones are mainly made up of ooids and superficial ooids, with subsidiary amounts of rounded shell debris, abundant peloids,

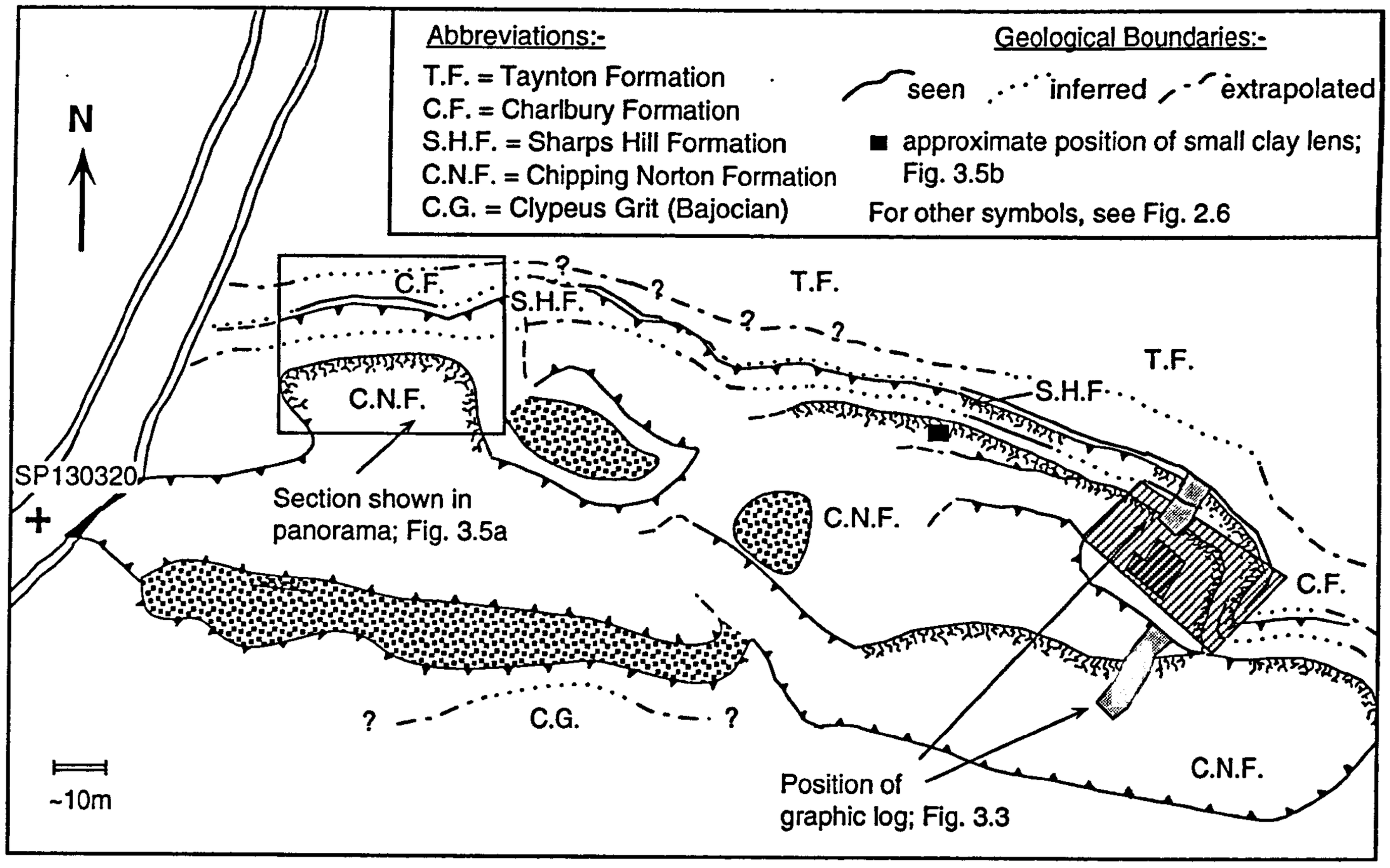


Figure 3.4. Sketch plan of the quarry floor, topography and small-scale geology after Vaughan (1989). The excavation site, the position of the graphic log (Fig. 3.3), and the section and subsidiary clay lens shown in Fig. 3.5. are also indicated. North (N) is shown.

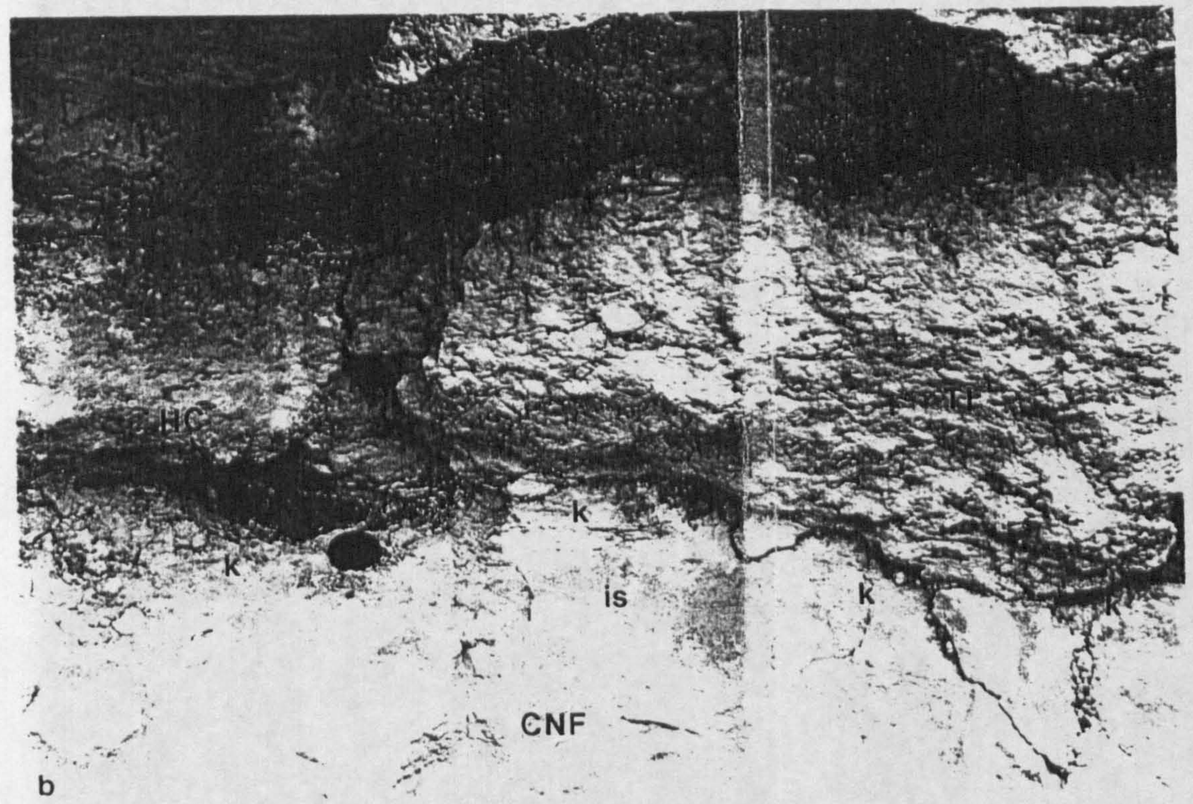


Figure 3.5. (a) Panoramic view of blasted face at Hornsleasow Quarry, about 150 metres west of the excavated clay lens. The palaeokarstic surface (k) can be traced throughout the face in the Chipping Norton Formation (CNF), although the clays are much less persistent. (b) A small pocket of clay (HC) found some 50 metres west of the main site overlying the ferric iron-stained (i.s.) palaeokarst (k). Overlying the lens is the transgressive sand unit (TL) which is 1-1.5 metres thick and can be traced along much of the quarry section, where the clays pinch out. North (N) is indicated, other abbreviations as Fig. 2.4.

composite grains, and homiolithic pebbles. Varying amounts of terrigenous material, in the form of quartz grains also occurs (Skerl, 1925, Horton *et al.*, 1987 and section 3.5.1). There are also thin beds of shelly marls (calcilutites) and some thin non-calcareous mudstones within the sequence. The mudstones and much of the terrigenous quartz in the formation, tend to occur more frequently to the east of the Moreton axis, where silica can make up to 30% of the rock (Sellwood & McKerrow, 1974). The limestones in this area are largely well-cemented by a coarse drusy-sparite cement (Skerl, 1925 and section 3.5.2) and are generally oosparites or oosparite-derivatives (Folk, 1959, 1962; Table C2, Appendix C). In places this cement has been partially decalcified to leave a limonite-stained friable sandy deposit, particularly noticeable in the field sections around Great Tew in northeast Oxfordshire (pers. obs.).

In the field, the limestones of the Chipping Norton Formation are well-bedded and locally form good freestones (Horton *et al.*, 1987). They show large scale tabular beds, or planar and trough cross-beds, with sharp, often erosional bases. Weathering produces more flaggy textures which highlight small-scale cross stratification and localised grading within the beds. Surfaces of individual cross beds may be ripple-laminated. Clay-filled burrows, plant debris and thin clay partings are also fairly abundant within individual horizons. The Chipping Norton Formation oolites are usually cream or pale buff in colour, but weathered sections tend to be highly iron-stained from dissolution of the ferroan calcite and can be anything from pale orange to bright greenish-yellow (Appendix C). Many of the sections of Chipping Norton Formation limestones show large-scale cambering joints (e.g. Fig. 4.1) and hand specimens can be riddled with post-burial cracks and non-ferroan calcitic veining.

3.5. Petrographic analysis of Chipping Norton Formation limestone surrounding the lens

A petrographic study was undertaken upon hand specimens of Chipping Norton Formation limestones which I removed from the study area in 1992. The exact sampling positions are marked upon Fig. C.1 (Appendix C3), and this diagram shows the relation between the specimens, and the palaeokarst and overlying soil horizons. In all, two blocks were removed from below the lens and two from the overlying strata. Each specimen was marked and numbered in the field before removal, to enable orientation during petrographic preparation. The specimens were prepared for study and description in several aspects, including cut and polished sections, stained polished surfaces and acetate peels, and petrographic thin sections (see Appendices C1-2).

The hand specimens, and the cut and polished blocks taken from them, were examined in conjunction with stained acetate peels of the polished surfaces, to facilitate recognition of large scale textural and mineralogical features of the rock type. These included the general bedding, lamination and cleavage relationships of the rock, an approximation of overall grainsize distribution and appraisal of any large clasts or allochems. The latter distinction is based upon a simple classification of limestones

proposed by Tucker (1981) which splits carbonate rocks into three main categories, calcirudite, calcarenite and calcilutite (for definition, Appendix C3). Also at this point any colour variation was noted and an assessment was made of the degree of weathering of the rock. This included a rough appraisal of how well cemented the rock was (Appendix C3). The most important petrological features of the unaltered Chipping Norton limestones in thin section and stained acetate peels are described in sections 3.5.1-3.5.8, and listed in full for each specimen in Appendix C3. The petrographic descriptions of the Hornsleasow limestones are split into two groups, those which occur below the clay lens (specimens HQCN1 and HQCN1': sections 3.5.1-3.5.4 and Fig. 3.6) and those which occur above the clay lens (specimens HQCN5 and HQCN6: sections 3.5.5-3.5.8 and Fig. 3.8). Their diagenetic histories are considered separately too (sections 3.6.1 and 3.6.2).

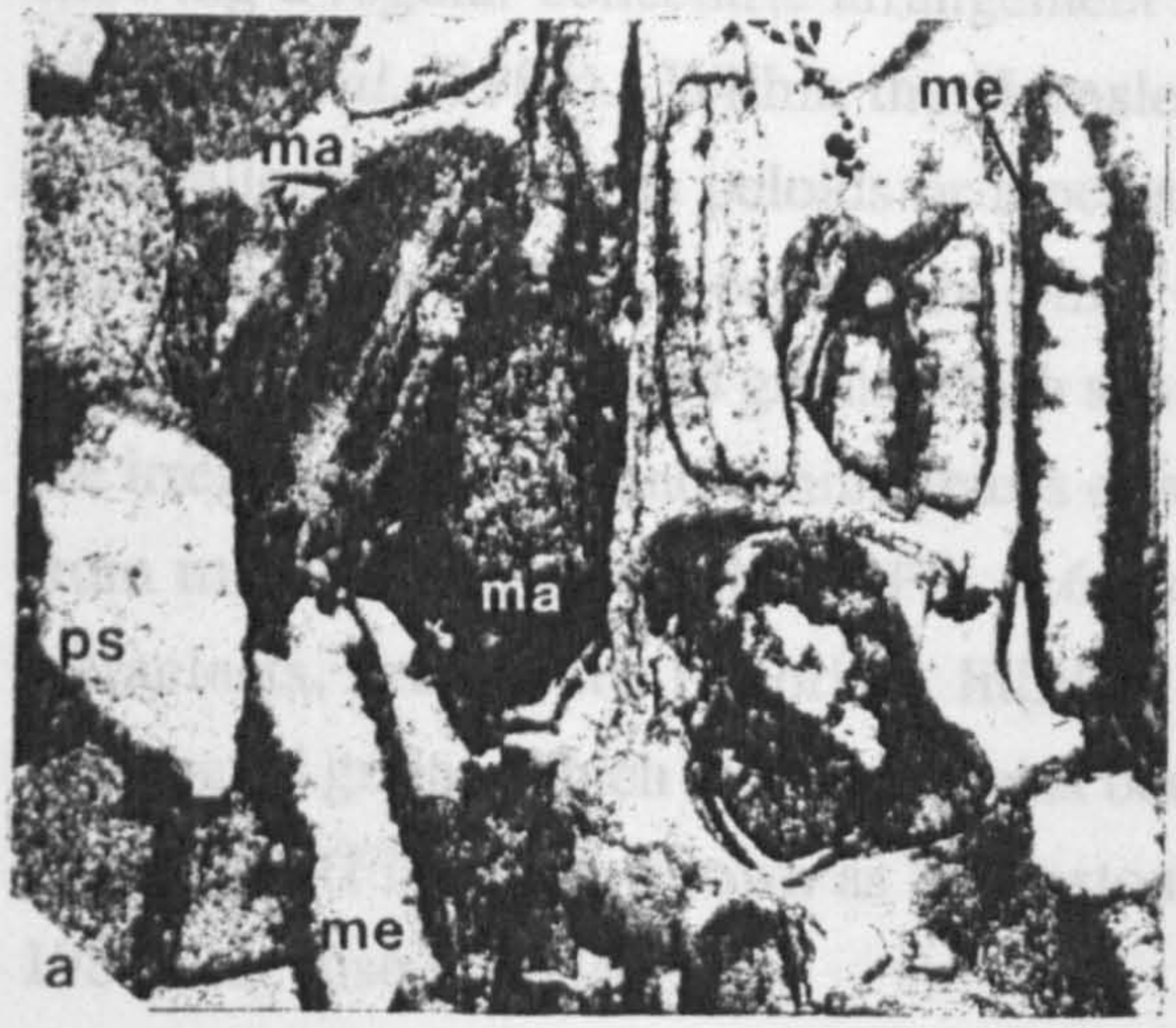
Generally, in hand specimen the Chipping Norton Formation limestones surrounding the clay lens are typical cross-bedded equigranular calcarenites composed of fine oolites, shell debris and peloids (Figs. 3.6 & 3.8). In detail, they show fine centimetre scale planar or gently inclined cross-lamination, and may contain thin laterally inconsistent lenses of coarser shell debris. There are occasionally larger beds of marly shell-fragmental limestones (Fig. 3.3), for example specimen HQCN6 (a calcilutite (Fig. 3.8d-f) which was taken from one of these beds above the lens at Hornsleasow (Fig. C1)) is composed of rare coarse biodebris (Fig. 3.8d) and ooliths (Fig. 3.8e) set within a matrix composed of carbonate mud or 'micrite'. This specimen is not well-cemented and quite friable. The calcarenites and calciruditic shell bands within them, are well cemented by a sparite cement (Figs. 3.6b,e,f & 3.8a-c). When fresh they are pale cream to buff-grey in colour, but weather to an iron stained warm yellow-buff colour. Contemporary weathering has caused a moderate amount of decalcification of the sparry cement, which is accompanied by rusty limonitic staining of the rock and causes grains to be dislodged (Fig. 3.8c). Some of the hand specimens, in particular HQCN1', exhibit coarse calcite veining and are pervaded with fine cracks.

3.5.1. Composition and textural analysis: Chipping Norton Limestones below the lens

The thin sections were examined in both natural daylight (in conjunction with the acetate peels) and under the petrographic microscope (Fig. 3.6). The former helps to recognise any distinct layering or lamination, the general grainsize distribution and sorting within the thin section before a description is attempted. Under the microscope the limestones were classified according to the two schemes for carbonate rocks proposed and developed by Folk (1959, 1962) and Dunham (1962). These classification schemes are based upon the proportions in the rock of allochems, cement and matrix.

Allochems are the large grains within the rock. They include bioclasts or skeletal grains, dominated in the Hornsleasow limestones by bivalves, in particular oyster material (Fig. 3.6a,c,f) and gastropods (Fig. 3.6e), as well as sparse brachiopods (Fig.

Figure 3.6. Photomicrographs of the Chipping Norton Formation unaltered limestone petrographic thin sections: (a) section HQCN1-1, plane polarised light; (b) HQCN1-1, cross-polarised light; (c) HQCN1-2, cross-polarised light; (d) HQCN1-2, cross-polarised light; (e) HQCN1', cross polarised light; (f) HQCN1', cross-polarised light. Field of view 2.3mm, except for (b) were field of view is 0.5mm. All sections were photographed under blue-filtered light. Abbreviations: (1) for allochems, bivalves 'bv', oysters 'oy', gastropods 'gp', brachiopods, 'bp', crinoids 'cn', echinoids 'ec', ooids 'oo', peloids 'pl', micritised allochems 'ma', micritised envelopes 'me', composite grains 'cp', intraclasts 'in', quartz 'q', opaques 'op'; (2) for matrix, micrite 'm'; (3) for cement, radial-fibrous calcite 'rf', drusy ferroan sparite 'ds', equant sparite 'eq', siderite 'si'; (4) for diagenetic features, pressure solution 'ps', fragmented allochems 'fg', squashed allochems 'sq', pseudo-unaxial cross 'ux'.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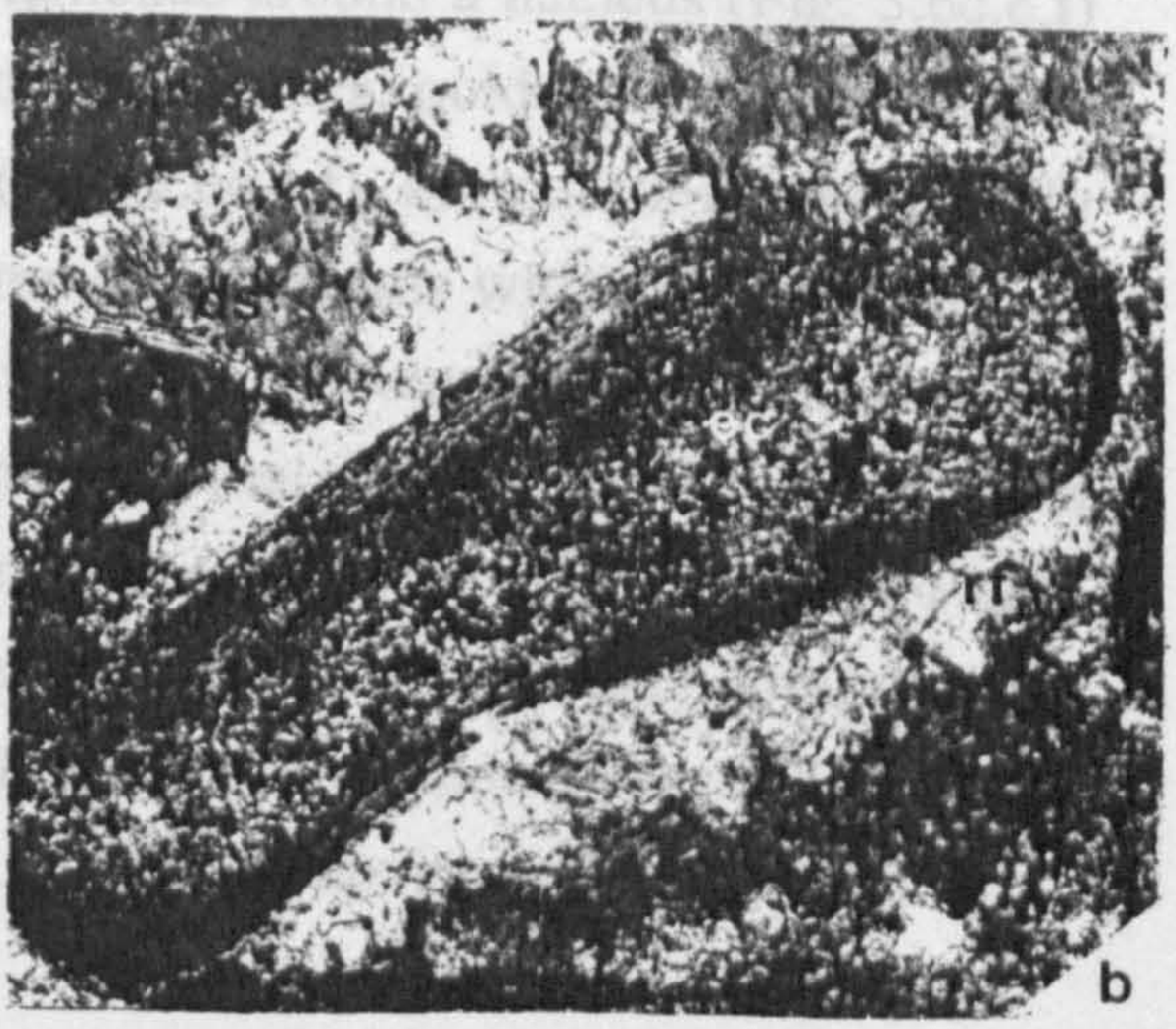
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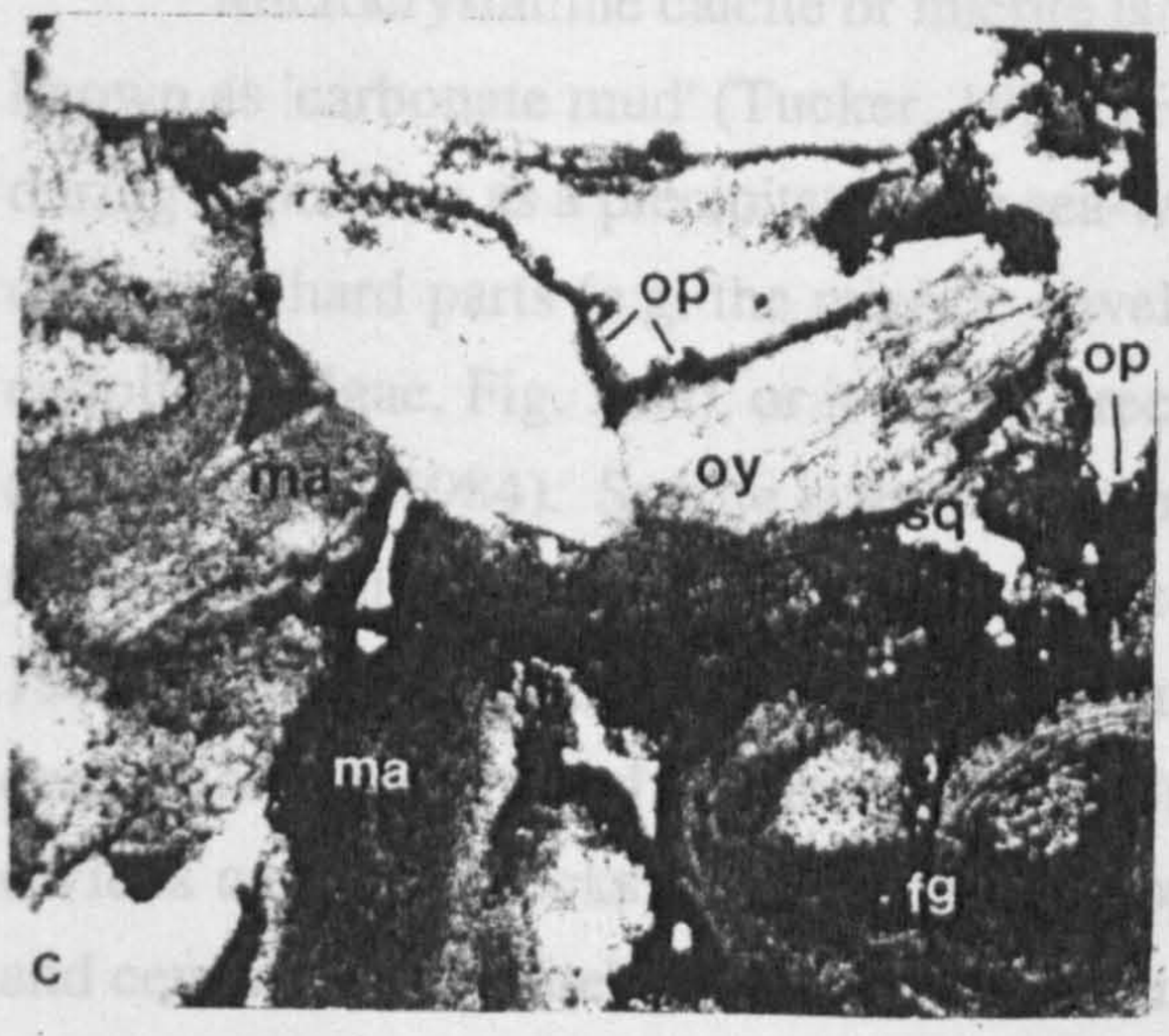
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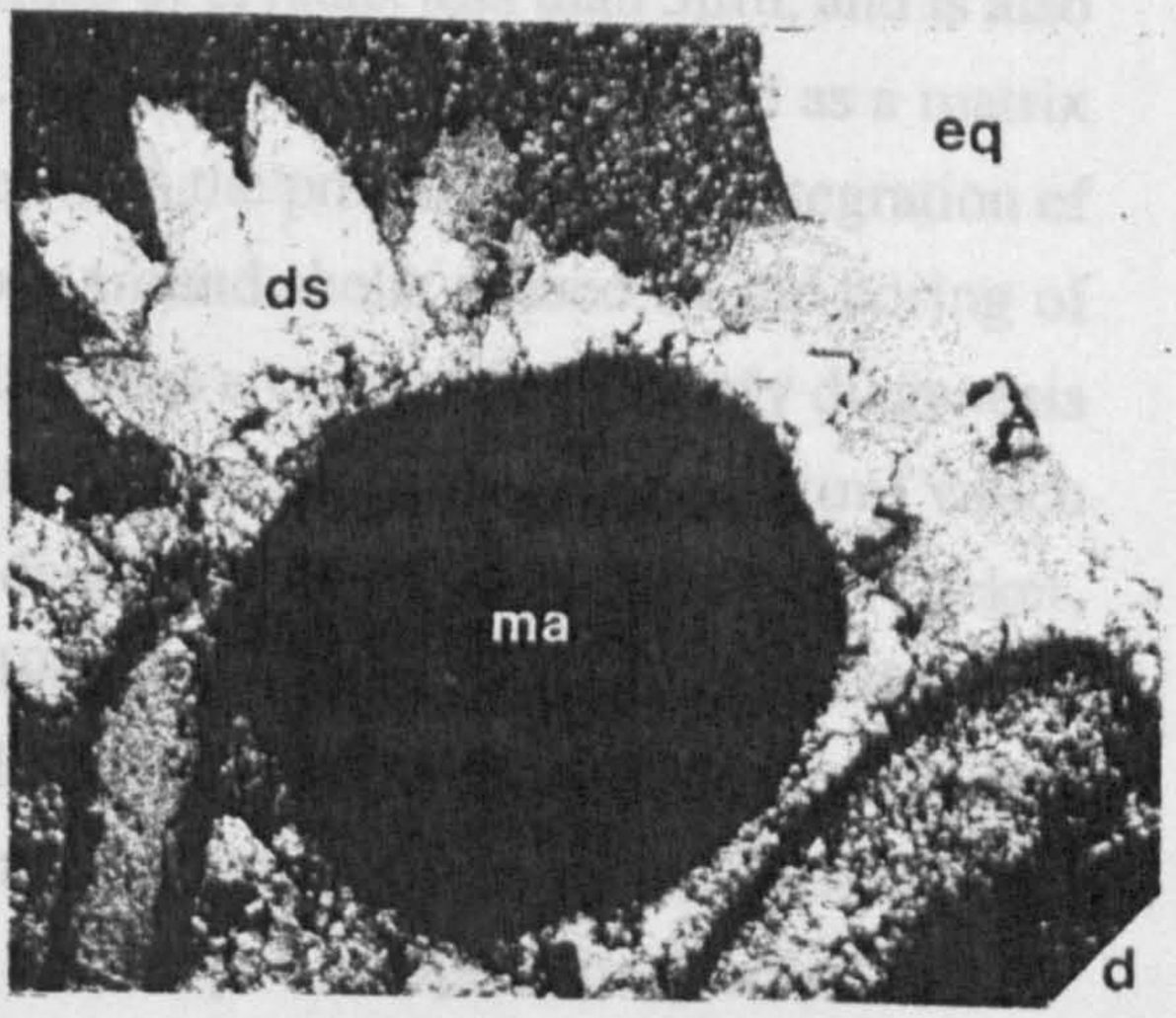
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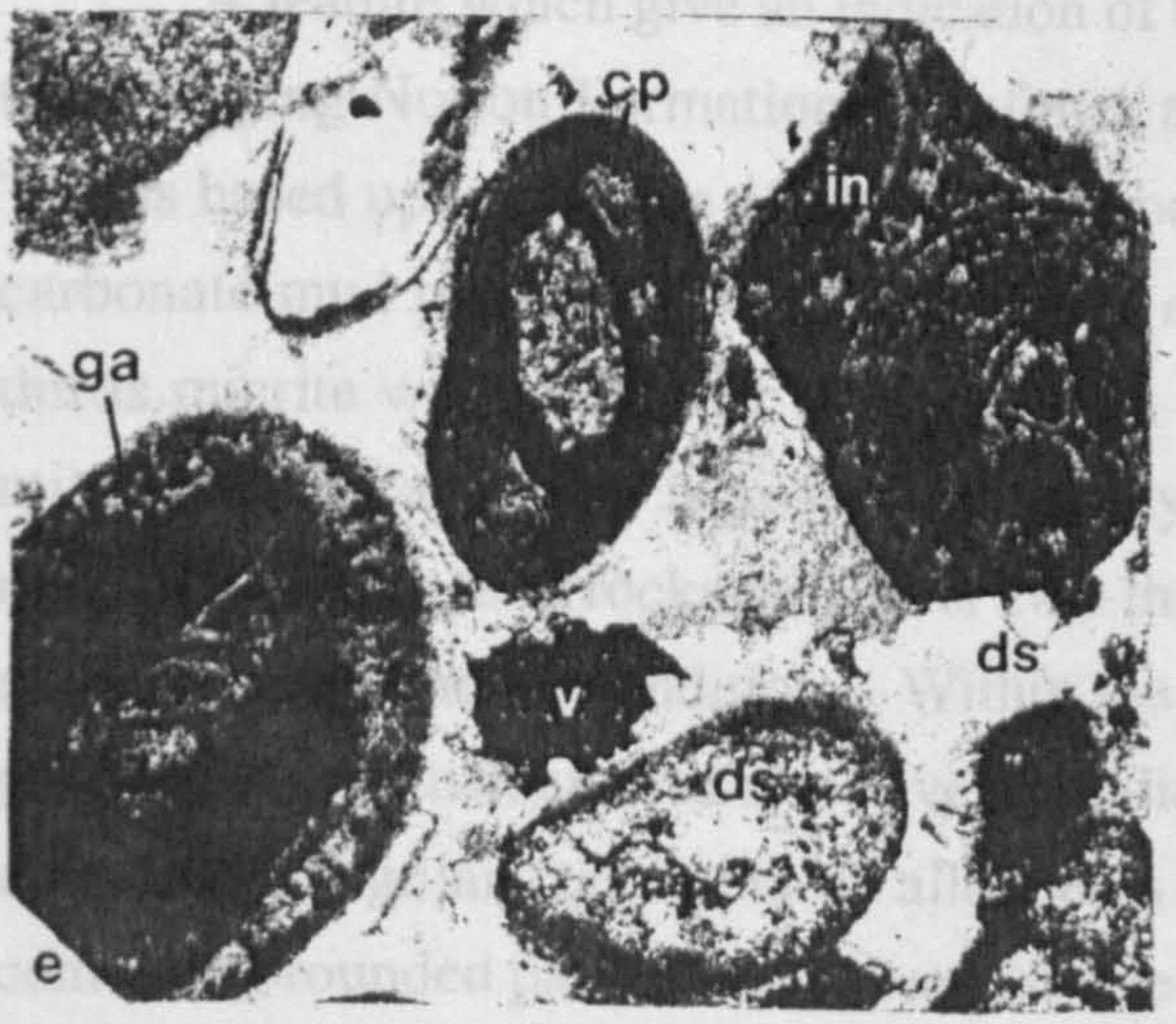
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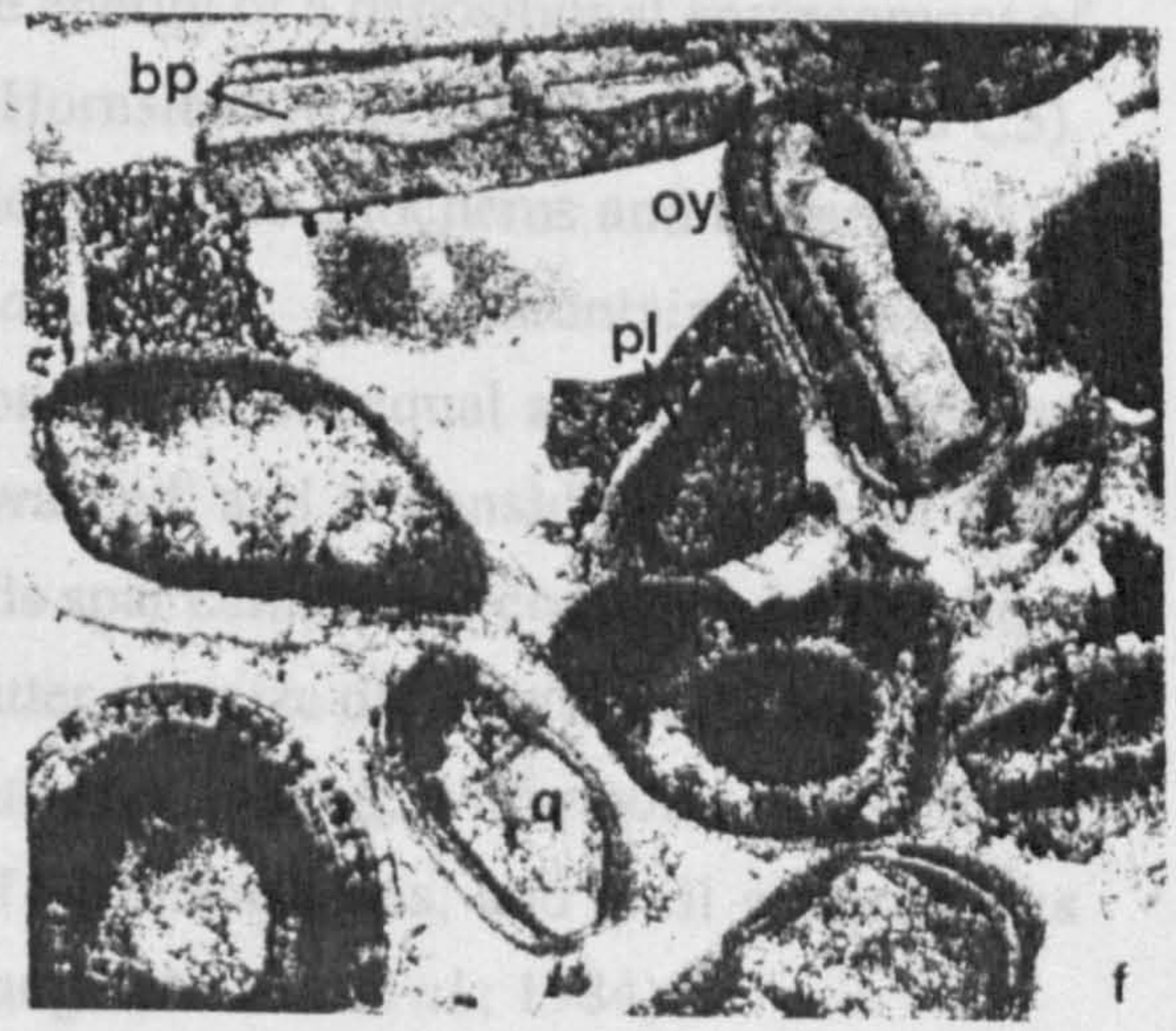
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3.6f), and echinoderms (Fig. 3.6a,b,d-f). Non-organic allochems include ooids or ooliths, which are the spherical and ellipsoidal grains of less than 2mm in diameter, showing a regular concentric arrangement of lamellae around a nucleus (Fig. 3.6c,e,f) (Adams *et al.*, 1984). Within the Hornsleasow limestones, ooids commonly enclose other allochems such as peloids or bioclasts (Fig. 3.6c,e,f), but quartz is also a fairly common nucleus (Fig. 3.6e,f). The other common allochems within the sections are peloids, entirely micritised grains which show no internal structure. Most of these grains are irregular in shape, and exhibit traces of a ghost internal structure and have originated from micritisation of bioclasts (Fig. 3.6c) or ooliths (Fig. 3.6d) (Adams *et al.*, 1984). Intraclasts, which are reworked lithified sediment (Fig. 3.6e), and composite or aggregated grains, which are small grains bound together by organic matter and cemented by micrite (Fig. 3.6f), known as 'grapestones' in modern environments (Adams *et al.*, 1984), are also abundant.

Microcrystalline calcite or micrite is formed of crystals less than 5µm, and is also known as 'carbonate mud' (Tucker, 1981) (Fig. 3.6a,c). It can be introduced as a matrix during deposition as a precipitate from sea-water or as the product of the disintegration of carbonate hard parts (e.g. the micritic envelopes around shells caused by the boring of endolithic algae, Fig. 3.6a), or it can be precipitated as a cement during early diagenesis (Adams *et al.*, 1984). Sparite is the coarse crystalline calcite (diameter over 5µm) which forms pore-filling cement and is usually a later diagenetic feature (Fig. 3.6b,d-f) (Tucker, 1981). The classification scheme of Dunham (1962) classifies limestones on the basis of depositional texture and is summarised in Table C1. The classification scheme of Folk divides carbonate rocks based upon the proportions of allochems, matrix (i.e. micrite) and cement (i.e. sparite) and hence, is a scheme based upon composition (Table C2).

A feature which give an indication of the energy of a depositional environment of the Chipping Norton Formation limestones at Hornsleasow is their sorting (Table C3). This is based upon degrees of mechanical rounding of the allochems and the amount of carbonate mud present in the slide (Adams *et al.*, 1984). A rock containing over two-thirds micrite was laid down in low-energy conditions, subequal amounts of spar and micrite within a section is known as 'poorly washed' and is considered to be of low-moderate energy, and rocks with over two-thirds spar cement are considered to be from moderate-high energy conditions. Within the latter, the size distribution and rounding of grains is also considered to also give an indication of sorting; with poorly sorted rocks containing angular fragments of allochems of differing sizes, and well sorted rocks containing rounded particles of sub-equal size-range (Adams *et al.*, 1984).

In terms of Dunham's classification, the majority of the limestones sampled from below the lens can be described as grainstones (Appendix C3; Fig. 3.6b,d-f). The shelly bands within the calcarenites are texturally rudstones. Carbonate mud has generally been removed from the rocks, but is present in small intergrain pores in amounts up to 15%

within the calcarenites (Fig.3.6a,c). Therefore they can be said to be 'poorly washed' packstones in places.

In terms of composition, the limestones from below the lens are cemented by a drusy spar cement composed of slightly ferroan calcite and are 'sparites' (Fig. 3.6a-f). Ooids, bioclasts and peloids are the most common allochems within the limestones and are sub-equal in proportions. The sections contain between 5-20% ooids (including grapestones and superficial ooliths), between 10-20% peloids and between 10-20% bioclasts (Appendix C3; Fig. 3.6). Other allochems make up about 5-10% of the rock, whilst non-carbonate grains are extremely rare. Quartz grains are fairly common as oolith interiors in specimen HQCN1', making up about 15-20% of all ooids in this section, in HQCN1 quartz is almost absent. Non-ooid quartz is much rarer and makes up only about 1-2% of the rocks by composition. All quartz grains appear to be terrigenous and were probably derived from the landmasses to the north of the area.

Opaques are rare (2-5% by composition) and are represented by diffuse 'smudges' of brownish-black or yellow-brown limonite (Fig.3.6c), tiny spherical grains of goethite (dark reddish-brown) or hematite (dark red) (Fig.3.6b,c,e,f) and black opaques which are probably pyrolusite or in some cases specks of carbon (Fig. 3.6a). Limonite 'stains' occur throughout the sections and have been noticed elsewhere in the formation (Skerl, 1925), but most opaques are concentrated within the weathering rinds around micritic envelopes on carbonate grains (Fig. 3.6a-e), within the laminae of ooids and within stylolitic sutures along which there has been dissolution of the sparite cement. Some allochems, and in particular micritised ooids and peloids, contain microscopic inclusions of limonite and spherules of goethite and hematite (Fig. 3.6b,c). The presence and origin of the opaque grains are considered in section 3.5.3.

3.5.2. Cementation and neomorphism: Chipping Norton Limestones below the lens

The morphology and mineralogy of carbonate cement crystals within a limestone can give much information about the environment of cementation and diagenetic history of the rock (section 3.6). The principal diagenetic processes are cementation, neomorphism and the effects of compaction after burial (Tucker and Wright, 1990). Carbonate rocks may encounter two major classes of diagenetic conditions as they are lithified, which are the 'vadose' zone, where pores are partially filled with water and partially with air, and the 'phreatic' zone, where pores are wholly water-filled (Sellwood, 1986) (Fig. 3.7). These zones can be further subdivided into marine and non-marine, and with active, moderate or stagnant circulation (Longman, 1980; Heckel, 1983) and each is characterised by particular diagenetic processes and cements (Fig. 3.7). For instance in the water-saturated marine phreatic zone there is an active zone where the sediments are affected by free-circulating waters caused by wave and current action, tidal movements and loss of CO_2 by degassing or photosynthesis (Fig. 3.7). In this zone carbonate cements composed of aragonite or high magnesium calcite are rapidly precipitated in the void

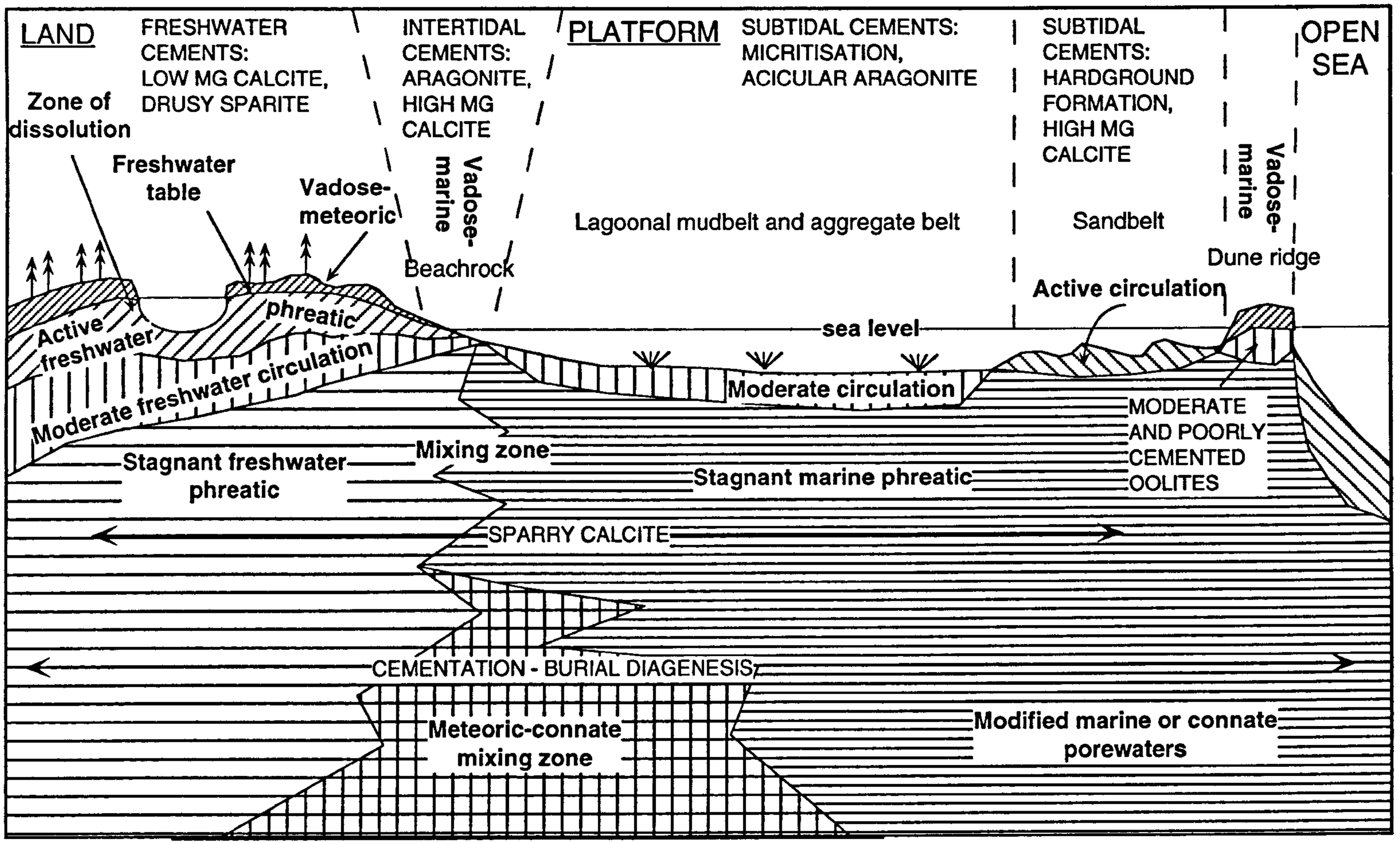


Figure 3.7. Schematic cross-section of shallow-marine carbonate platform and raised terrestrial platform environment, with zones of cementation, burial and diagenetic processes indicated (after Tucker, 1981, Reading, 1986).

space between grains (Heckel, 1983). However in the stagnant marine phreatic zone below the surface (Fig. 3.7), there is very little free circulation of interstitial water and so cementation is limited to small pores and borings (where it is precipitated as micritic cement: Heckel, 1983). Hence the grains in the former limestones are much less likely to suffer compaction with burial. A further zone shown on Fig. 3.7 is the deep-burial or 'connate' zone (Longman, 1980). The various cements and their formation in the zones shown in Fig. 3.7 are considered further in section 3.6.

In order to understand the diagenetic histories of the limestones from below and above the lens (section 3.6), the thin sections and acetate peels were studied under natural daylight in order to recognise zones of different cementation (Appendix C3). This was facilitated by the application of the Dickson (1966) method of staining which was employed to distinguish different carbonate cements (Appendix C2) within the acetate peels. Chemical staining works by producing a precipitate on the surface of a particular mineral. The resulting colour stains produced by using the combined Dickson solution for each type of carbonate cement usually found in limestones are:

Aragonite	Pink
Calcite	Pink
Ferroan calcite (>1% $FeCO_3$)	Mauve-Royal blue
Dolomite	Unstained
Ferroan dolomite (>1% $FeCO_3$)	Turquoise

Potassium ferricyanide is very sensitive to the detection of iron in carbonates and will stain cements with as little as 1% iron carbonate within their matrix. With increasing iron content the colour varies from mauve -> purple -> royal blue in calcite cements and from pale turquoise -> deep turquoise in dolomitic cements (Dickson, 1966). Other useful tools in determining the composition of carbonate cements and from them the diagenetic history of the limestones, are studying the thin sections with cathodoluminescence (CDL) and ascertaining their bulk and trace element geochemistry (Tucker and Wright, 1990). CDL causes the manganese present in the calcite cement lattice to luminesce. Iron in the lattice acts as an inhibitor to the luminescence, and so the degree of luminescence is related to the ratio of iron to manganese in the cements (Frank *et al.*, 1982; Fairchild, 1983, 1988; Sun, 1990; Wright and Tucker, 1990). Therefore the varying degrees of luminescence within a section can be used to determine chemical composition of the cements and from that in which order and under what conditions they were most likely to have been precipitated (i.e. their diagenetic history: Sun, 1990).

A CDL and geochemical analysis of the Hornsleasow limestones was not carried out because of the time span of the project. I feel that this has limited the use and interpretation of the petrographic descriptions of the Chipping Norton Formation

limestones above and below the lens, and the karstic limestones at Hornsleasow (Chapter 4) and hope that in the future I can carry out such a study.

In nearly all the sections four distinct types of cement were noticed within the Hornsleasow sections, in variable proportions (Appendix C3):

(a) Pore-filling micrite or calcite microspar. Micrite or dark cryptocrystalline matrix may occasionally form a cement in ancient limestones, however its distinction from sedimentary carbonate mud is not easy to establish (Adams *et al.*, 1984). It occurs mostly as a cavity lining or coating to grains (Fig. 3.6a,c) and may also form bridges between grains (e.g. Fig. 3.6c shows dark brown micrite bridge between two heavily green micritised bioclasts on the left-hand side of the photograph). It is also possible that the micrite may have been deposited as carbonate mud with the grains as a wackestone or packstone. This could have been partially lithified and then subjected to removal of unlithified material by pore-water erosion (Adams *et al.*, 1984).

(b) Radial-fibrous pore-lining calcite. This is rare within the Hornsleasow sections occurring in only the better preserved parts of the specimens. A thin radial-fibrous fringe of calcite is observed upon a superficial oolite grain, with an echinoid fragment as the core, in Fig. 3.6b. In this particular case the crystals are elongate bladed in form and the fringe is less than 100μm thick. In some cases, the fringe shows unit extinction in cross-polars with the allochem it surrounds. This type of optical continuity is called a 'syntaxial overgrowth' (Adams *et al.*, 1984) and occurs when a cement nucleates upon a substrate of the same mineralogy. Most of the radial-fibrous fringes are isopachous suggesting that they were formed within the phreatic zone (Fig. 3.7).

(c) Drusy mosaic pore-filling slightly ferroan calcite spar. The main pore-filling cement within the grainstones below the lens at Hornsleasow is a slightly ferroan (stained pale mauve) calcite sparite or drusy spar (Fig. 3.6a-f). The general features of this cement are its interstitial location between grains and within skeletal grains; that it is a clear cement with few inclusions (Fig. 3.6a-f) and that it occurs in a 'drusy' fabric i.e. where the crystals become more coarse from the outside of a pore into the interior (Fig. 3.6d,e). The drusy fabric can be entirely void-filling (Fig. 3.6b,f) or partially pore-filling with void spaces ('v' on Fig. 3.6e). In up to 30% of cases it is also seen to have infilled dissolved skeletal grains which exhibit ghost outlines maintained by their micritic envelopes (Fig. 3.6e,f). Rare large poikilotopic crystals (i.e. crystals which enclose smaller crystals) are also seen.

(d) Equant calcite vein-spar. Large vug-filling and vein non-ferroan calcite (stained pink) is seen both in polished and thin sections of the Hornsleasow limestones. The Chipping Norton limestones below the lens are often riddled with cracks and veins (specimens HQCN1-2 and HQCN1': Appendix C3). In many cases the veins transect grains and dissolution is seen along the grain surface. The veins can also cut through the drusy cements in some cases (Appendix C3)

The various environments of formation for each of these cements seen in the Hornsleasow limestones and their diagenetic history is considered in detail in section 3.6.1.

3.5.3 Preservation of allochems and cement: Chipping Norton Limestones below the lens

The preservation and mineralogy of the grains and the matrix within the limestones can also yield much information upon the original conditions of deposition and any post-depositional alteration and diagenesis of the rocks. This is particularly relevant for the Hornsleasow limestones which have undergone karstification and sub-aerial weathering in Bathonian times (Chapter 4), but was also studied for the unaltered limestones below and above (section 3.5.7) the clay lens.

The recrystallisation of a mineral to a polymorphic form of itself is known as 'neomorphism' (Adams *et al.*, 1984) and in carbonates, the two most common neomorphic transformations are aragonite and high magnesium calcite to low magnesium calcite (Tucker & Wright, 1990). Both these reactions take place in the presence of water and occur by solution and reprecipitation across a reaction front. It is also known as 'calcitisation' (Adams *et al.*, 1984, Tucker, 1990), particularly when it forms an *in situ* replacement of skeletal carbonate grains. Neomorphism includes replacement of acicular aragonitic radial-fibrous cements by fibrous calcite and also the alteration of micritic matrix to a coarser 'microspar' or 'pseudospar', a process known as aggrading neomorphism (Tucker, 1981). The latter occurs within calcilutites and is thought to occur within meteoric pore-waters.

Bioclasts and ooids within the Hornsleasow limestones are sometimes preserved by calcitisation. Calcitisation takes places through the neomorphic transformation of aragonite to calcite, without an intervening void phase, and it is usually quite rare (e.g. Sun, 1990). However, in specimens HQCN1-1 and HQCN1' a few (<5%) originally aragonitic bivalves appear to have been calcitised, these are now composed of sparry slightly ferroan calcite, but contain lines of inclusions which cut across crystal boundaries and appear to trace out the original foliated structures (e.g. Fig. 3.7b). Most ooids if unmicritised show a radial fabric, as well as the usual concentric arrangement of lamellae (e.g. Figs. 3.6f). This radial structure is quite common in ancient ooliths (see examples in Tucker, 1981; Adams *et al.*, 1985; Tucker & Wright, 1990), although their formation is a point of contention (section 3.6.1).

Around 45-50% of the allochems within the Chipping Norton Formation grainstones below the clay lens are partially or wholly altered to micrite, by a process known as 'micritisation' (Bathurst, 1966, 1975). Micritisation is caused by the biochemical erosion and alteration of the skeletal carbonate to micrite by an endolithic boring coccoid blue-green alga or cyanophytea (Tucker, 1981), which forms an irregular dark micritic 'envelope' around the edge of the grain. Micritic envelopes may also show deeper borings into the interior of the allochem, for example Fig. 3.6a shows a bivalve

fragment with a well-developed dark green micritic envelope, in which borings penetrate about 10μm into the shell. Borings are usually infilled with micrite after vacation by the alga. Further degradation of the allochem can lead to a partially or totally micritised grain, which may or may not exhibit 'ghost' structures (Fig. 3.6a,c-f). These are known as peloids, but can be separated from pelletal peloids by their irregular shape and remnant structure. The abundance of micritic envelopes and micritised bioclasts within the limestones is considered in section 3.6.1.

Around 5-10% (rising to between 30-40% in specimens HQCN1-2 and HQCN1') of the bioclasts within the Chipping Norton limestones below the lens have been dissolved away to leave a void which was later partially or wholly infilled with a drusy mosaic sparite cement. In these cases, the original bioclasts (gastropods and bivalves, e.g. Fig. 3.6e) were composed of aragonite. They are now infilled with a slightly ferroan calcite cement (stained pale mauve) which exhibits none of the original internal structure of the shells, although the original shape of the bioclast was maintained during dissolution by either an early cement fringe or the micritic envelope (Fig. 3.6e).

Opaques are rare in the grainstones, but partial dissolution of the ferroan carbonate cements (including siderite) to leave a residue of ferric hydroxide and aggregated iron oxides has also taken place in some of the sections. This is particularly noticeable in the outer more ferroan-rich lamellae of ooids (Figs. 3.6b) where oxidation of the ferroan iron has left a black opaque residue of aggregated limonite. Although some of this opaque residue may be organic material or pyrite. Pyrite occurs locally within organic-rich ooid laminae and replaces micrite envelopes in other Jurassic limestone sequences (Sun, 1990). Sparse granules of goethite and limonite also occur within the drusy sparite cement (Fig. 3.6a).

Grain-overpacking is fairly common in these sections (Fig. 3.6a,c) and pressure solution during burial is indicated by the presence of compacted and deformed allochems and grain-to-grain dissolution contacts in the limestones. Pressure-solution takes place when a sediment is compacted under a substantial load, and in limestones it is the carbonate which is usually dissolved, to leave a residue of insoluble material such as clay, ferric oxyhydrites and quartz grains concentrated in thin seams known as 'stylolites' (Adams *et al.*, 1984). In all the non-karstic Hornsleasow sections, stylolitic sutures are rare (section 3.5.7), but in the limestones underlying the lens softer grains, such as peloids and some oolites, may be deformed and squashed into one another or into harder grains (Fig. 3.6a,c).

3.5.4. Porosity: Chipping Norton Limestones below the lens

An evaluation of both type and amount of porosity within the limestone was also examined. Porosity can either be a primary depositional feature of the rock or secondary, forming from dissolution as a result of diagenesis or shattering of the rock by tectonic processes. There are three main types of porosity seen in limestones (Adams *et al.*,

1984), and these are 'fabric-selective', 'non-fabric-selective' and 'partially-fabric-selective'. These are split up into several sub-classes, the most common of which are described below:

(a) Fabric-selective porosity can be formed by primary or secondary processes and includes: interparticle (primary pores between grains or matrix removed by secondary dissolution); intercrystal (cement removed by secondary dissolution); intraparticle (selective dissolution of parts of allochem during diagenesis); mouldic (allochems dissolved away during diagenesis); fenestral (secondary dissolution along laminae or bedding planes); and shelter (primary depositional hollows beneath curved allochems e.g. bivalve shells. It can also occur within chambered fossils when partially infilled within sediment these form geopetal or way-up structures).

(b) Non-fabric-selective is usually secondary in origin and includes: fracture (cracks formed from compaction and tectonic stresses); channel (late diagenetic dissolution along cracks); and vug (discrete secondary dissolutional hollows).

(c) Partially-fabric-selective porosity can be primary or secondary in formation and includes: breccia (physical breaking apart of rock may follow non-fabric selective cracks or fabric-selective fenestrae); boring (a primary feature, which in shells is fabric-selective, otherwise is not); burrows; and shrinkage (irregular pores caused by primary desiccation of the unlithified sediment. Within micrites these are also known as 'birdseyes'; Tucker, 1981).

Initially porosity is quite high and should be about 50% in sand-grade carbonate sediments (Tucker, 1981). However, this primary porosity is lost during cementation, compaction and pressure-solution. Hence primary porosity of the well-cemented Hornsleasow grainstones is consequently very low (1-2%). However, porosity can be gained through solution and fracturing, and the sections show up to 5% vug and channel porosity, although in most cases this too has been infilled by sparite cements. Within weathered surfaces dissolution of the cement and grain loss is up to about 10-15%.

3.5.5. Composition and textural analysis: Chipping Norton Limestones above the lens

Compositionally, the limestones sampled from above the lens at Hornsleasow are grainstones (HQCN5) or wackestones (HQCN6). Allochems make up between 40-65% of the rocks by composition. Bioclasts are reasonably abundant, making up between 5-15% of the rock, and are mainly bivalve (Fig. 3.8b,d), and echinoderm (Fig. 3.8b,d) fragments. Although other forms such as gastropods, brachiopods, bryozoans and foraminifera are also found in the sections. Ooids are abundant (15-20%) (Fig. 3.8a-f) and commonly enclose other allochems such as peloids or bioclasts (Fig. 3.8a-f), or quartz (Fig. 3.8c). Fig. 3.8e shows a tiny shark tooth (0.5mm in length) forming the interior of an ooid. Peloids tend to be pelletal in shape and would have originated as faecal pellets of animals (Fig. 3.8b,e) (Tucker, 1981), although some (Fig. 3.8a,c) are more irregular and exhibit traces of a ghost internal structure and would have originated

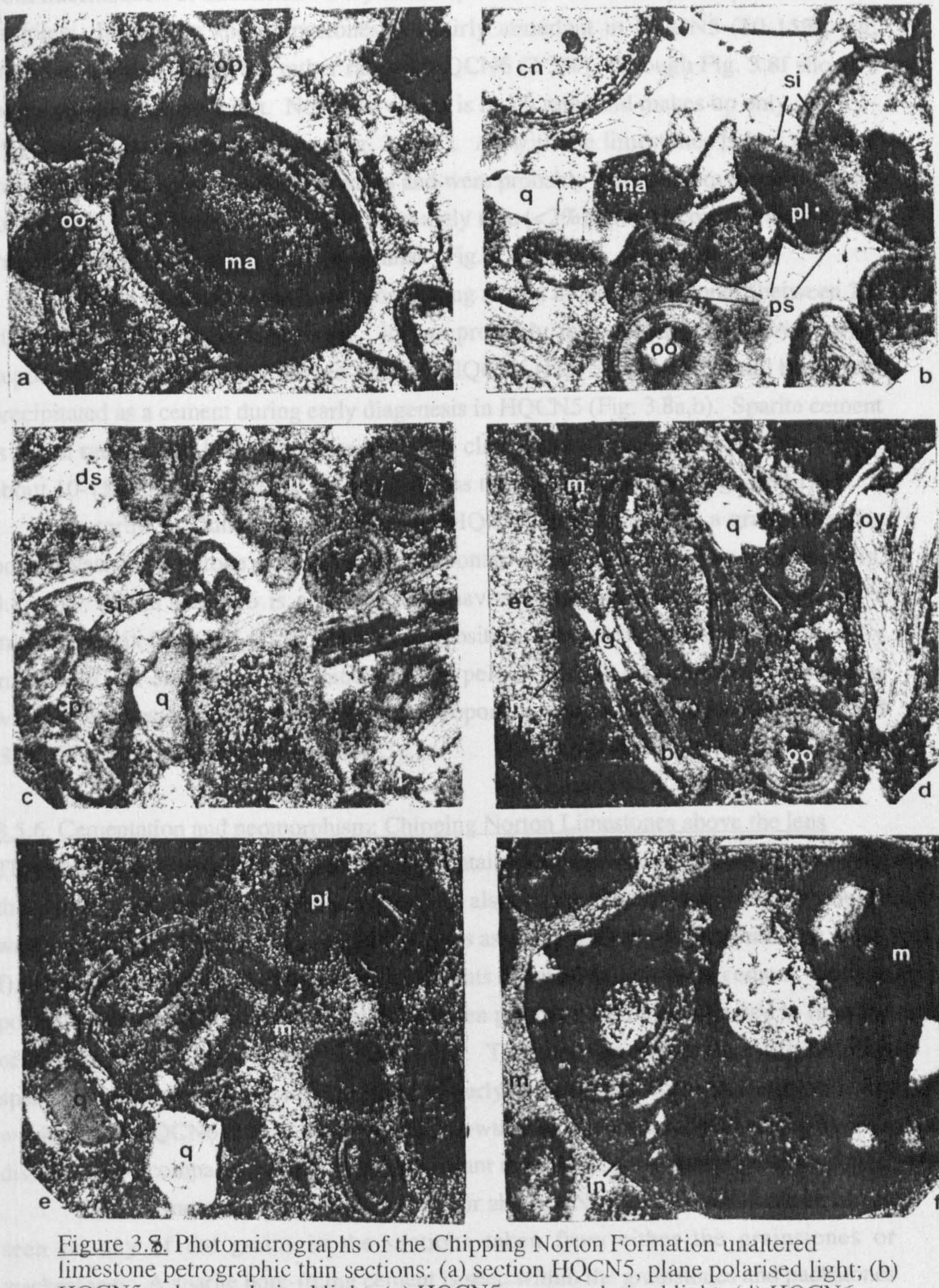


Figure 3.8. Photomicrographs of the Chipping Norton Formation unaltered limestone petrographic thin sections: (a) section HQCN5, plane polarised light; (b) HQCN5, plane polarised light; (c) HQCN5, cross-polarised light; (d) HQCN6, cross-polarised light; (e) HQCN6, cross polarised light; (f) HQCN6 cross-polarised light. Field of view 2.3mm. All sections were photographed under blue-filtered light. Abbreviations as in Fig. 3.6.

from micritisation of allochems. They make up between 10-15% by composition in the samples. Intraclasts and grapestones are fairly abundant in HQCN5 (10-15%; e.g. grapestone in Fig. 3.8c), but rather rarer in HQCN6 (2-5%), although Fig. 3.8f shows a large and rounded intraclast. Non-ooid quartz is much rarer and makes up only about 1-2% of the rocks by composition (Fig. 3.8b,e). As with the limestones below the lens, the quartz grains appear to be terrigenous and were probably derived from the landmasses to the north of the area. Opaques are extremely rare (<2%) and are concentrated within the weathering rinds or within ooid laminae (Fig. 3.8a,b,d-f).

Micrite is fairly common as a pore-filling matrix in these limestones (between 20-50% of the rock) (Fig. 3.8a,b,d-f). It was probably introduced as a matrix during deposition as a precipitate from sea-water in HQCN6 (Fig. 3.8d-f) or it could have been precipitated as a cement during early diagenesis in HQCN5 (Fig. 3.8a,b). Sparite cement is much rarer in the limestones overlying the clay lens at Hornsleasow, and makes up about 10-15% of HQCN5 (Fig. 3.8a-c) and less than 5% of HQCN6 (Fig. 3.8d-f).

In terms of Dunham's classification, HQCN5 ranges between a grainstone to a poorly washed packstone with up to 25% carbonate mud as a matrix (Appendix C3; Fig. 3.8a-c). Whilst HQCN6 is a wackestone having sparse allochems supported by a micritic matrix (Fig.3.8d-f). In terms of composition, the limestones are compositionally 'micrites' (Fig. 3.8d-f). Ooids, bioclasts and peloids are the most common allochems within the limestones and are sub-equal in proportions. HQCN5 is an oobiomicrite and HQCN6 is a oopelmicrite (Appendix C3).

3.5.6. Cementation and neomorphism: Chipping Norton Limestones above the lens

The sections taken from above the clay lens contain much more interstitial micrite in them than those below the lens (section 3.5.2). It is also possible that the most of this micrite was deposited as carbonate mud with the grains as a wackestone or packstone (Fig. 3.8d-f). In sample HQCN5 there is subequal amounts of micrite and sparite cement, and it is possible that the carbonate mud could have been partially lithified, subjected to removal of unlithified material by pore-water erosion. The resulting voids being filled by the sparite cement (Adams *et al.*, 1984). An early micritic cement is indicated for the wackestone, HQCN6 (Fig. 3.8d-f), which shows little evidence of grain breakage and dissolution by compaction and contains abundant rounded lithified intraclasts (Fig. 3.8f).

Other cements in these rocks are rare or absent. No radial-fibrous cements were seen on any of the grains in the sections taken from either the grainstones or wackestones. A sparite pore-filling cement is seen within the grainstones and packstones (Fig. 3.8a-c) and in some pores within the wackestones (Fig. 3.8e) and this is a ferroan calcite (stained mauve), anhedral drusy spar (Fig. 3.8a-c). It is a relatively clear cement with few inclusions (Fig. 3.8b,c). In the rare occasions (between 5-10% in HQCN5 and <5% in HQCN6) that the dissolved bioclasts have been infilled with drusy spar (e.g. Fig. 3.8b), the cement is not completely pore-filling. Vein calcite was not seen in these

specimens, although it was observed at outcrop in some of the oosparites overlying the clay lens. The formation of these cements and the diagenetic history of the limestones above the lens is considered in detail in section 3.6.2.

3.5.7 Preservation of allochems and cement: Chipping Norton Limestones above the lens

The calcilutite, specimen HQCN6, exhibits patches of neomorphic microspar within the micritic matrix (Fig. 3.8d,e), which may have formed during aggrading neomorphism. In both specimens taken from above the clay lens, some bioclasts (5-10%), mainly aragonitic bivalves, appear to have been calcitised and are now composed of sparry ferroan calcite, with original foliated structures preserved (e.g. Fig. 3.8b-d). Most ooids in these sections have been calcitised, preserving their original concentric structure and showing a radial fabric and an associated pseudo-uniaxial cross (Fig. 3.7c-d) which is seen in cross-polarised light.

Most allochems show the development of a micritic envelope around the edge of the grain and around 60-70% of the allochems are partially or wholly altered to micrite (micritised allochems 'ma' in Fig. 3.8a,b,e,f). Dissolution of the sparite intergrain and intragrain cements is extremely rare in these sections. Opaques are also extremely uncommon (<2%) and are only seen in organic-rich lamellae of ooids (Fig. 3.8a,b).

Stylolitic sutures are rare in the limestones overlying the clay lens, but grain-to-grain pressure-solution at points where allochems come into contact are quite common (Fig. 3.8a-c). Also softer grains are deformed and some brittle allochems have been broken (Fig. 3.7d).

3.5.8. Porosity: Chipping Norton Limestones above the lens

Although in a lime mud sediment there is up to 80% initial porosity (Tucker, 1981), the Hornsleasow packstones and wackestones from above the lens are all largely well-cemented and primary porosity is consequently very low (<5%). The sections show up to 5% vug and channel porosity, although in most cases this too has been infilled by later sparite cements and in some cases (e.g. HQCN6) fracture porosity may have occurred during the preparation of the thin sections (Appendix C3). Within weathered surfaces dissolution of the cement and grain loss is up to about 10-15%.

3.6. Depositional environment and diagenetic history of the Chipping Norton Formation limestones in the quarry and in the northern Cotswolds

In general, the sediments of the Chipping Norton Formation to the west of the Moreton axial structure were generally deposited in shallow-water, high-energy conditions with unstable substrates. This is supported by the low diversity and fragmented nature of the bioclasts within the limestones (section 8.1). The fauna would appear to be largely composed of shallow marine epifaunal and shallow infaunal forms, with a predominance of oysters and echinoderms (Skerl, 1925). Ooliths usually form in agitated and warm

waters, by inorganic precipitation of aragonite from sea water supersaturated in calcium carbonate. These are frequently moved as sand-waves, dunes and ripples at the edge of the marine platform. Oolite shoals are seen fringing the Bahaman Platform (Hine, 1977). However, calcitic ooids can also form by bacterial activity within organic grain coatings creating a microenvironment for carbonate precipitation, in quiet water lagoonal and intertidal conditions (Tucker, 1981; Tucker & Wright, 1990).

Sedimentation rates appear to have been high, although the appearance in the sequence of several hardgrounds and intensely bioturbated beds, as well as the common occurrence in petrographic sections of reworked, composite and micritised grains (Skerl, 1925), suggest that at times there may have been substrate erosion and colonisation. This is also indicated at Condicote (SP146270), six kilometres to the south of Hornsleasow, where storm deposited rhythmically bedded quartzitic oolites alternate with bioturbated oolites (Sellwood & McKerrow, 1974). The presence of calcilutites, abundant faecal pellets and aggregated grains in the formation in the northern Cotswolds suggest deposition within shallow subtidal areas, as in modern circumstances they are commonly found within lagoonal areas associated with algal mats upon the Bahaman platform (Tucker, 1981). The limestones were probably deposited in conditions analogous to those which prevail within the modern-day Bahaman shallow water carbonate platform.

There appears to have been some terrigenous input from the north and northeast, as the limestones contain some detrital quartz, which most likely originated from contemporary weathering and reworking of Palaeozoic sediments in the Midlands region (Chapter 6). The formation thins toward the Moreton axis from over twelve metres at Hornsleasow to around four metres on the swell (Sellwood & McKerrow, 1974) and this may have formed a local high upon the platform at that time (Chapter 6).

In the next sections I shall attempt to interpret the features seen in the petrographic analyses of the Chipping Norton Formation limestones which occur above and below the clay lens at Hornsleasow. A depositional and diagenetic model for each of these sequences is also proposed.

3.6.1. Depositional and diagenetic history of limestones deposited below the clay lens: a regressive sequence

The sediments below the clay lens at Hornsleasow are dominantly coarse grained. They comprise mainly oolitic grainstones which exhibit cross-bedding at outcrop. The macrofaunal component of the oolitic grainstones which occur below the lens at Hornsleasow is rather impoverished, comprising mainly oyster fragments and gastropods, and any burrows tend to be vertical, suggesting that substrates were unstable. These large scale features suggest that the limestones were deposited in an extremely shallow subtidal environment with agitated waters and restricted fauna. The terrigenous input (wood and quartz grains) suggests a close proximity to land areas. These limestones occur between half metre and one metre below a highly weathered,

palaeokarstic horizon and soil (Chapters 4-5) and so were deposited in a regressive sequence.

In detail the grainstones are composed of oolitic, bioclastic and peloidal grains. Modern ooids usually form in tropical, normal marine, shallow subtidal (under two metres depth) agitated waters, such the high-energy Bahaman Banks (Hine, 1977). In this environment the ooids are moved around as ooid shoals, although localised cementation of the seafloor can give rise to intraclasts, hardgrounds and cemented crusts (Tucker & Wright, 1990). In the modern environment, ooids are composed of aragonite and exhibit strong concentric lamellae formed by a tangential arrangement of the acicular crystals, which often contain inclusions of organic material.

Ooids can also form within low energy, restricted lagoonal and tidal marine environments (Tucker and Wright, 1990). They are much more rare, and are composed of low magnesium calcite. They may show a radial arrangement of the acicular calcite crystals and are frequently broken along the grain of this microfabric (Tucker, 1981).

The Hornsleasow ooids show both concentric and radial fabrics. Most authors (e.g. Shearman *et al.*, 1970, Adams *et al.*, 1985; Sun, 1990) favour the hypothesis that the radial fabric is purely diagenetic and is formed during slow neomorphism of the original aragonitic ooid to low magnesium calcite in the marine phreatic environment which also preserves original concentric textures. Their reasoning is because aragonitic ooids are much more abundant in modern circumstances than calcitic ooids, and indeed, rare radial structures have been seen in ooids composed of aragonite or high magnesium calcite precipitated in modern marine environments (Tucker & Wright, 1990). However, Wilkinson & Landing (1978) have suggested that many Jurassic ooids were probably calcitic and that the radial fabric is a primary feature. This might reflect a different sea-water chemistry in the Jurassic than at present (Tucker & Wright, 1990). Hence, if the radial structure seen within the Chipping Norton Formation ooids is primary rather than diagenetic, this would indicate that they were formed in a fairly quiet tidal-dominated palaeoenvironment.

The preservation of the allochems is also suggestive of a shallow platform environment of deposition. Almost all of the allochems in the Hornsleasow limestones bear well-developed micritic envelopes and over half of bioclastic grains are partially or entirely micritised. Micritised bioclasts and ooids are a common component of modern Bahaman Platform carbonates and algal borings may be cited to indicate that the limestone was deposited within the photic zone, but other organisms such as fungi, bivalves and sponges bore bioclasts, and transportation from the photic zone into deeper water has to be taken into consideration. Micritisation of grains is most prevalent in quiet, low energy locations where there is little sediment movement (Tucker & Wright, 1990), it can also develop in intertidal environments were 'beachrock' is being cemented (see below).

Micrite cement is much rarer in these sections, although aggregated grapestones are cemented by micrite. In modern sediments micrite forms cement fringes around

allochems in marine pore-waters, which are usually of equal thickness or 'isopachous' (e.g. Fig. 3.6a,c), indicating that they were precipitated below the water-table within the marine 'phreatic' zone (Fig. 3.7) (Longman, 1980; Heckel, 1983). Unequal or 'meniscus' fringes indicate that the pores were only partially filled with water, and therefore, that the cement was precipitated within the 'vadose' zone (Fig. 3.7) (Sellwood, 1986), but such fringes are not observed in the Hornsleasow limestones. In modern carbonate environments such as the Bahaman Platform, the microcrystalline cement is precipitated as a cryptocrystalline aragonitic mud within the less agitated parts of the platform such as in lagoons and upon tidal flats (Tucker, 1981; Tucker and Wright, 1990). It is thought to be formed from a combination of inorganic precipitation of aragonite within the sea-water and the disintegration of calcareous algae (Neumann & Land, 1967) and larger carbonate allochems by mechanical, physio-chemical and biochemical means (Tucker, 1981). Some allochems in the Hornsleasow limestones show mechanical breakage and evidence of dissolution along their edges (Figs. 3.6a,c).

Such large scale micritisation of grains and the formation of micrite-cemented grapestones in the Hornsleasow limestones is strongly suggestive that the oolite shoal environment in which the sediments were laid down stabilised, allowing microboring biological and mechanical disintegration of the grains and limited early micritic lithification between grains in their original depositional environment. Grapestones commonly form in low to moderate energy, extremely shallow (usually less than 5m deep) environments, were currents are sufficient to remove lime mud and silt, but not sand sized particles (Tucker and Wright, 1990). Further cementation would encourage hardground development (see below). No such hardground is seen in the section at Hornsleasow below the lens (Fig. 3.3), although elsewhere in the Chipping Norton Formation they are reasonably developed. The karstic horizon underlying the clay lens may have originally been a hardground horizon (Chapter 4).

The isopachous pore-lining calcite cements which surrounds the grains in the grainstones below the lens post-date micritic envelope formation, but are near-surface early cementation phenomenon, which probably took place in the oolite shoal environment in the active zone of the marine phreatic environment (Longman, 1980). The precipitation of such cements in modern-day environments leads to the formation of hardground horizons or surface crusts, which develop just below the sediment-water interface were grains are not being moved about continuously, but water is being flushed through the sediment (Tucker and Wright, 1990). If exposed, these cemented portions may become bored and colonised by hardground biota of cementing epifaunal and shallow infaunal animals. Such encrusted horizons are common in the Great Oolite Group limestones and present elsewhere in the Chipping Norton Formation.

In modern day marine-platform pore waters, precipitation of aragonite occurs close to the sediment-water interface to form fringes of needle-like, acicular crystals (Tucker and Wright, 1990). Precipitation of high magnesium calcite can also occur

within reef or sandbelt habitats (marine or freshwater phreatic zones) and these form more prismatic, columnar or bladed crystals (Sun, 1990). The fringes seen in the Hornsleasow specimens could originally have been aragonitic, in which case inversion to low magnesium calcite during diagenesis could have destroyed the original acicular habit of the crystals. Alternatively, they may have been precipitated as high magnesium calcite, later reverting to low magnesium calcite by neomorphism (Land, 1971; Longman, 1980). The precipitation of high magnesium calcite cement can be synsedimentary or form during early diagenesis in the marine environment (Sellwood, 1986), and the occurrence of synsedimentary hardgrounds within the Chipping Norton Formation sequence would appear to confirm this hypothesis (cf. Purser, 1969). Early cementation also indicated by the presence of intraclasts.

Absence of meniscus or pendant fabric suggests that the primary cementation took place in water filled pore-space and was not precipitated in the marine vadose zone along the shoreface (Longman, 1980). However, early pendant or meniscus cements can be masked by later cement growth and neomorphism of the earlier precipitates, and isopachous cements can form in the lower parts of the intertidal zone (Tucker and Wright, 1990). In the marine vadose zone which occurs at the shoreline or in subaerially exposed shoals ('beachrock' and 'dune ridge' environments on Fig. 3.7), the pore-space just below the beach surface is well aerated and allows rapid evaporation and CO_2-degassing from the marine porewaters. This causes cements to precipitate rapidly as fibrous or acicular aragonite and micritic high magnesium calcite to form beachrock (Tucker and Wright, 1990). Beachrock is continuously broken up into intraclasts (up to cobble and boulder sized) which are rounded and re-cemented into more beachrock. The formation and features of beachrock are considered in more detail in Chapter 4, when discussing the early formation of the karstic surface at Hornsleasow.

To summarise then, the Chipping Norton Formation limestones underlying the lens at Hornsleasow Quarry have undergone primary cementation on the sea-floor or within the marine substrate, either by acicular aragonitic or fibrous high magnesium calcite. An alternative origin for cement fringes was described by Talbot (1971) and Sun (1990) who documented widespread cement fringes in the Jurassic limestones of southern England which appeared to have been precipitated in shallow marine, slightly reducing phreatic environments as low magnesium calcite. Some of these were apparently slightly ferroan in composition reflecting the reducing nature of the surrounding pore waters, but as Sun (1990) explained this geochemical feature may still be diagenetic after neomorphism of high magnesium calcite or aragonite.

After this primary sea-floor cementation of the limestones, dissolution or neomorphism of aragonite seems to have occurred. The fringe cements in the Hornsleasow limestones are present only in the smallest pore-spaces and only occur on non-leached grains, which suggests that many may have been removed by dissolution before sparite precipitation. This would indicate that the cement was originally composed

of aragonite or high magnesium calcite rather than low magnesium calcite. That any fringes survive at all in the rocks suggests that neomorphism to low magnesium calcite took place and was reasonably rapid. Longman (1980) suggested that rapid neomorphism takes place in the freshwater phreatic environment, but in a stagnant zone (Fig. 3.7) where the pore-waters are saturated in dissolved carbonate. It is related to the permeability of the sediment, as with low permeability the solution-reprecipitation front is rather narrow allowing re-precipitation without a significant void space developing, resulting in the preservation of the original fringe cement texture (Chowdhury, 1982). The variation of preservation of the cement in the Hornsleasow carbonates reflects the differences in permeability, with the fringes being preserved in only the smallest pore spaces. As the limestones occur below a palaeokarstic surface the neomorphism would have taken place under the influence of Bathonian meteoric conditions. Chowdhury (1982) described a similar case where primary aragonitic or high magnesium calcite cements in the Osmington Oolite Formation (Upper Oxfordian) were precipitated in the marine phreatic zone and were neomorphosed to low magnesium calcite in the freshwater phreatic zone during contemporaneous uplift of the limestones in the Oxford area.

The calcitisation of aragonitic bioclasts may have taken place during this phase of diagenesis. Within the Chipping Norton Formation limestones, well-preserved (non-micritised) ooids exhibit the radial structure described above, and also a slight compositional zoning from inner lamellae of low magnesium calcite to the outer lamellae composed of ferroan calcite. This zoning could suggest that these ooids were also slowly calcitised during phreatic diagenesis. However, these ooids were probably neomorphosed from high magnesium calcite or aragonite in anoxic, reducing pore-waters (meteoric or marine) at some depth below the sediment water interface (Sun, 1990).

As well as neomorphism, leaching of primary aragonitic bioclasts in the Hornsleasow limestones is indicated by the presence of ghost micritic envelopes which trace out former allochems. Micritic envelopes and micritised grains are not leached. Leaching of aragonitic or high magnesium calcitic cements and allochems could have taken place in either the meteoric phreatic or vadose environment, where there was active circulation of carbonate-undersaturated waters (Gvirtzman & Friedman, 1977; Chowdhury, 1982; Heckel, 1983). This suggests that the limestones were being flushed through with freshwater, after moving into the meteoric environment (Fig. 3.7). Similar leaching has been observed in regressive sequences in the Corallian (Upper Oxfordian) succession of southern and central England, were an influx of meteoric water was caused by emergence of the limestones at the top of the sequence in the Oxfordshire shallows region (Chowdhury, 1982; Sun, 1990).

The early fibrous pore-lining fringe cements do not occur on the inside of collapsed micrite envelopes (cf. Sun, 1990), suggesting that leaching of aragonitic grains occurred after these early cements had been precipitated and in some cases this dissolution removed these early cements too. The void space left by the leached grains is

cemented by a drusy slightly ferroan sparite cement, indicating that leaching took place before spar cementation. The Hornsleasow drusy-replaced grains have not been compacted during the intervening time period between dissolution and sparite precipitation, which suggests that this interlude must have been quite short. However, grain overcrowding and the presence of deformed soft grains suggest a limited amount of compaction following the leaching of the early fibrous cements and prior to sparite precipitation. The intergranular sparite cement in the limestones has a coarser grainsize than that precipitated within leached grain pore space, but without geochemical or CDL studies it is difficult to know whether the cements are contemporaneous or if one was precipitated before the other.

Drusy spar is the main diagenetic cement formed within meteoric waters, where dissolution of the original skeletal carbonates in the form of aragonite and high magnesium calcite, and reprecipitation of low magnesium calcite occurs (Longman, 1980; Tucker, 1981). However, cementation in the marine environment cannot be fully ruled out for the sparry calcite, as in some Jurassic hardgrounds it has been precipitated in shallow marine sediments (Wilkinson *et al.*, 1985).

Meteoric cements are invariably of the drusy or equant crystalline form, because the fresh pore-waters have lower magnesium:calcium ratios than sea-water and magnesium is known to inhibit crystal growth (Tucker & Wright, 1990). Sparite precipitation in the limestones below the lens probably took place within the active meteoric phreatic zone, with free circulating oxygenated waters, rather than in the zone of precipitation in the meteoric vadose environment as no unequal or meniscus cements were observed (Heckel, 1983; Sun, 1990). Although equant and bladed low magnesium calcite spar can be precipitated in large pore-spaces in both environments. For instance Weiss and Wilkinson (1988) undertook a similar study of modern biosparites from a regressive sequence from the central Texas coast and concluded that the cements seen in the carbonates were precipitated from meteoric fluids during subaerial exposure, and indicated replacement of marine phreatic environment by meteoric waters, dissolving unstable grains and cements and precipitating carbonate cements in oxidising environment.. Low magnesium syntaxial overgrowths on echinoderm grains are also formed in the active saturated meteoric phreatic zone, but these occur very rarely in the Hornsleasow grainstones.

Usually, sparite cement laid down in the meteoric diagenetic environment is low magnesium calcite, mobile iron in vadose and active freshwater phreatic zones tends to form inclusions of iron hydroxides and oxides (Heckel, 1983). However, both the intergranular and intragranular sparite in the grainstones is mildly ferroan calcite, suggesting slightly reducing conditions of precipitation. If the cement was laid down as ferroan low magnesium calcite it is probably a late burial cement (Longman, 1980; Heckel, 1983) and was laid down in mixing zone or even connate zone waters (Fig. 3.7). This suggests that the limestones underlying the karstic surface at Hornsleasow were

largely uncemented (except for the early cement fringes) during exposure in the early Bathonian and that this cement was precipitated long after inundation and burial. That some of the grains are squashed into one another and the limestones show grain-overpacking might indicate that this is the case. Heckel (1983) described a similar diagenetic history for a regressive sequence in Pennsylvanian age carbonates, were re-submergence during the following transgressive cycle caused renewed precipitation of cements of progressively more ferroan contents in the void spaces. However, the sparite may have been precipitated as a high magnesium calcite cement in the marine or meteoric phreatic environment and picked up ferrous iron during later neomorphism in a reducing phreatic or even connate environment (cf. Chowdhury, 1982).

The overlying karstic limestones in the profile at Hornsleasow were exposed and possess strong meteoric vadose dissolution features (Chapter 4). The limestones described here are only half and one metre below the karst surface (Fig. C1) and were probably influenced by meteoric phreatic waters, although whether they ever entered the vadose environment is unclear. The high ferroan nature of the cement suggests that it was precipitated or neomorphosed in reducing conditions. However, the cement does contain tiny inclusions of iron oxyhydrites which could have been introduced in three ways. If the cement was precipitated before uplift, the inclusions may have been formed by oxidation of the ferrous iron in the calcite lattice during Bathonian vadose dissolution (Chapter 4), however if this were the case large vugs in the cement should occur (none were seen). The second method could be through introduction of iron particles into the uncemented sediments in the vadose zone during the Bathonian karstification and prior to sparite cementation (Foos, 1991: Chapter 4). The third method could be through modern day weathering and oxidation of the ferrous cements.

The equant low magnesium calcite spar cement which infills solution veins was probably precipitated during deep burial. It has been suggested that sparite can be precipitated during deep burial within connate (i.e. modified sea-water) and mixed connate-meteoric pore-waters (Longman, 1980). There are several sources of the carbonate within the deep burial environment: firstly, from the connate water; secondly, from pressure solution within the limestones; and finally, from solution of skeletal carbonate within marls and shales interbedded with the limestones (Tucker, 1981). Carbonate sediments which are cemented early within the vadose zone or active phreatic zone are less likely to be modified than those which undergo diagenesis in the deeper phreatic zone. The latter show a greater amount of grain alteration and dissolution porosity. Both the ferroan drusy spar and veining spar were precipitated after the majority of grain deformation and pressure-solution in the Hornsleasow limestones. However, stylolitic pressure solution appears to have been post-drusy formation and probably formed during deep burial.

Finally, uplift in recent times and dissolution occurred between the meteoric vadose and phreatic zones.

3.6.2. Depositional and diagenetic history of limestones deposited above the clay lens: transgressive sequence

In terms of composition the limestones above the clay lens at Hornsleasow tend to contain many more skeletal fragments, more peloids and much more interstitial micrite than those below the lens. Shell-fragmental and pelletal biomicrite beds are much more common in the section (Fig. 3.3.) and faunal diversity is higher in these beds. Bioturbation is more apparent in the section, with vertical, horizontal and inclined burrows occurring (Fig. 3.3). Directly overlying the clay lens is a bioclastic sand horizon (HQTR1 and HQTR1': Fig. C1) interpreted as a beach deposit (section 5.13). All these features suggest that the limestones above the clay lens were deposited in shallow, but gradually deepening subtidal conditions of a fairly low-energy shoreline. Similar features have been seen in modern sediments deposited during the slow drowning of exposed carbonate platforms, such as around Andros Island on the Bahaman platform (Tucker & Wright, 1990) and northwest Florida coastline (Hine *et al.*, 1988).

The depositional and diagenetic history of the limestones overlying the palaeokarstic surface and associated terrestrial deposits is as follows. The limestones were laid down in low energy shallow marine environments with intense micritisation of grains and micrite cementation in the active phreatic zone. The absence of any early pore-lining fringe cements is rather perplexing in these limestones, as they were deposited in shallow marine conditions, but is probably a function of the lower energy conditions in which the muddy sands were laid down. In more protected areas there is less water circulation through the phreatic environment, meaning porewaters are more stagnant and so, cementation is a localised phenomenon (Tucker and Wright, 1990). The presence of intraclasts, grapestones and some intergranular cement suggests that this was the case. About one metre above these limestones a hardground horizon is developed in similar bioclastic limestones, and has been encrusted by oysters and bored (Fig. 3.3). This would suggest that precipitation of high magnesium calcite and aragonite was taking place in these carbonates, sufficiently to caused stabilisation of the substrate.

Early pore-lining cements may have been leached from the sediments by undersaturated meteoric pore-waters in the phreatic environment before further cementation. Although the limestones above the lens at Hornsleasow were deposited in deepening marine conditions, following the inundation of the karstic surface and paleosols (Chapters 4 and 5) they were probably still influenced by meteoric waters. The strong overpacking and compaction of grains prior to sparite cementation suggests that these sediments were only weakly cemented during early burial.

As with the limestones underlying the lens at Hornsleasow, neomorphism and limited dissolution of skeletal aragonite has taken place. Dissolution of the grains may have been caused by meteoric water influences, but progressive burial of the sediments would have restricted the influence of meteoric water influx. As well as the

neomorphosed bioclasts, there are well-preserved ooids in the limestones which exhibit radial structures and a strong compositional zoning from inner lamellae of low magnesium calcite to the outer lamellae composed of ferroan calcite or even golden-brown siderite ($FeCO_3$, which exhibits bright birefringence colours in cross polars) (Fig. 3.8b,c). This zoning suggests that these ooids were slowly neomorphosed during diagenesis in the stagnant marine phreatic zone (Heckel, 1983).

The ferroan nature of the intergranular and intragranular sparite suggests that the cement was precipitated as low magnesium calcite. The pervasive distribution of strongly ferroan void-filling calcite spar in the Hornsleasow packstones and grainstones was probably cemented in low oxygen marine phreatic or deeper burial connate zone. The sparite cement which infilled the pores in other transgressive sediments such as those in the Osmington Oolite Formation (Upper Oxfordian) is also ferroan (Chowdhury, 1982), and modern biomicrites sampled from a transgressive sequence in the central Texas region were cemented in the presence of marine and marine-derived fluids, possibly mixed with connate fluids as they were buried (Weiss and Wilkinson, 1988).

Chapter Four Hornsleasow palaeokarst

4.1. Introduction to palaeokarsts

Limestone land areas form distinctive landscape features, such as cave passages, blind valleys, springs, underground streams and bare rock surfaces or 'limestone pavements'. On a smaller scale, the limestone surfaces may show dissolution runnels, closed depressions or 'dolines', and smooth planed surfaces (Trudgill, 1985). Collectively, limestone landforms are known as 'karst', named after the Kras area of Yugoslavia, where there is a well-developed limestone landscape (Trudgill, 1985). Karst is defined as the distinctive assemblage of landforms developed upon soluble bedrock (Jennings, 1971). The small-scale erosional features of subaerial karstic surfaces are known as 'lapies' or 'karren' (Trudgill, 1985).

The features of a limestone landscape are formed by the action of erosion and dissolution, and in particular, the dissolutional action of water upon the soluble calcium carbonate framework of the rock itself. Most karsts are formed by dissolution and removal of the solute under freshwater conditions, for example in fluvial regimes, by the action of groundwater or acidic porewaters within soils. However, coastal karsts also form under the combined actions of dissolution, biological and biochemical activities, and mechanical abrasion. Carbonate dissolution does not take place at an equal rate over the whole exposed surface of the rock, but may be concentrated along joints and fissures within the limestone. This arises as a result of enhanced water flow along such features, causing their enlargement. The process becomes particularly pronounced with the formation of underground cave passages. Other erosional and weathering processes are fairly important in the formation of karstic landscapes. These include the effects of frost (particularly in temperate and arctic regions), mechanical abrasion (by physical and organic means such as tree root activity), hydration, biological action and biochemical processes (Trudgill, 1985). These processes also effect less soluble rock types and are not restricted to limestone regions.

Many limestones are composed of fairly pure calcium carbonate, therefore once they have dissolved only a thin residual soil cover or 'rendzina' will remain. This soil cover is easily removed by downward transport into underground cavities and run-off to leave a bare rock surface (Trudgill, 1985), for example the limestone pavements of the Burren, County Clare, Eire. Limestone pavements which develop under thick soil cover are known as 'covered' karst or 'cryptokarst' (Jennings, 1971, 1985); their karren forms are often more subdued than those formed by direct dissolution of the bare rock. Surface drainage may be minimal on well-developed limestone scenery, for instance the dry river

valleys of Cheddar Gorge, Avon and the Castleton area of the southern Peak District, Yorkshire. In these areas water action usually takes place in underground streams and cave passages, although during major flooding events old valleys may carry water along them (Trudgill, 1985).

Karst landforms are a ubiquitous feature of all limestone land areas and are widespread throughout the world with well-developed karst present in arctic, temperate, tropical and arid conditions. Karst formation is not climate-controlled, although climatic conditions, such as precipitation levels and soil cover, may exert a control upon the scale of the dissolution features formed (Trudgill, 1985). Well known Holocene karstic landscapes have formed upon the indurated and well cemented Lower Carboniferous limestones exposed at the surface in the Ingleborough-Malham district of the northern Pennines of Yorkshire, the Burren in County Clare, and in parts of south Wales and the Mendips. In these regions, well developed karren, cave and dry river passages are found, and have been described by many authors (e.g. Sweeting, 1972). The younger Mesozoic limestones of Great Britain and in particular, the friable and poorly cemented oolites of the Middle Jurassic exhibit less noticeable, small-scale karstic features. However, oolitic limestones in other parts of the world and in particular the tropics, have been severely altered by karstification, for example, the Tertiary and Holocene limestones of the Bahaman Platform and Florida Bay area, U.S.A. (e.g. Hine *et al.*, 1988).

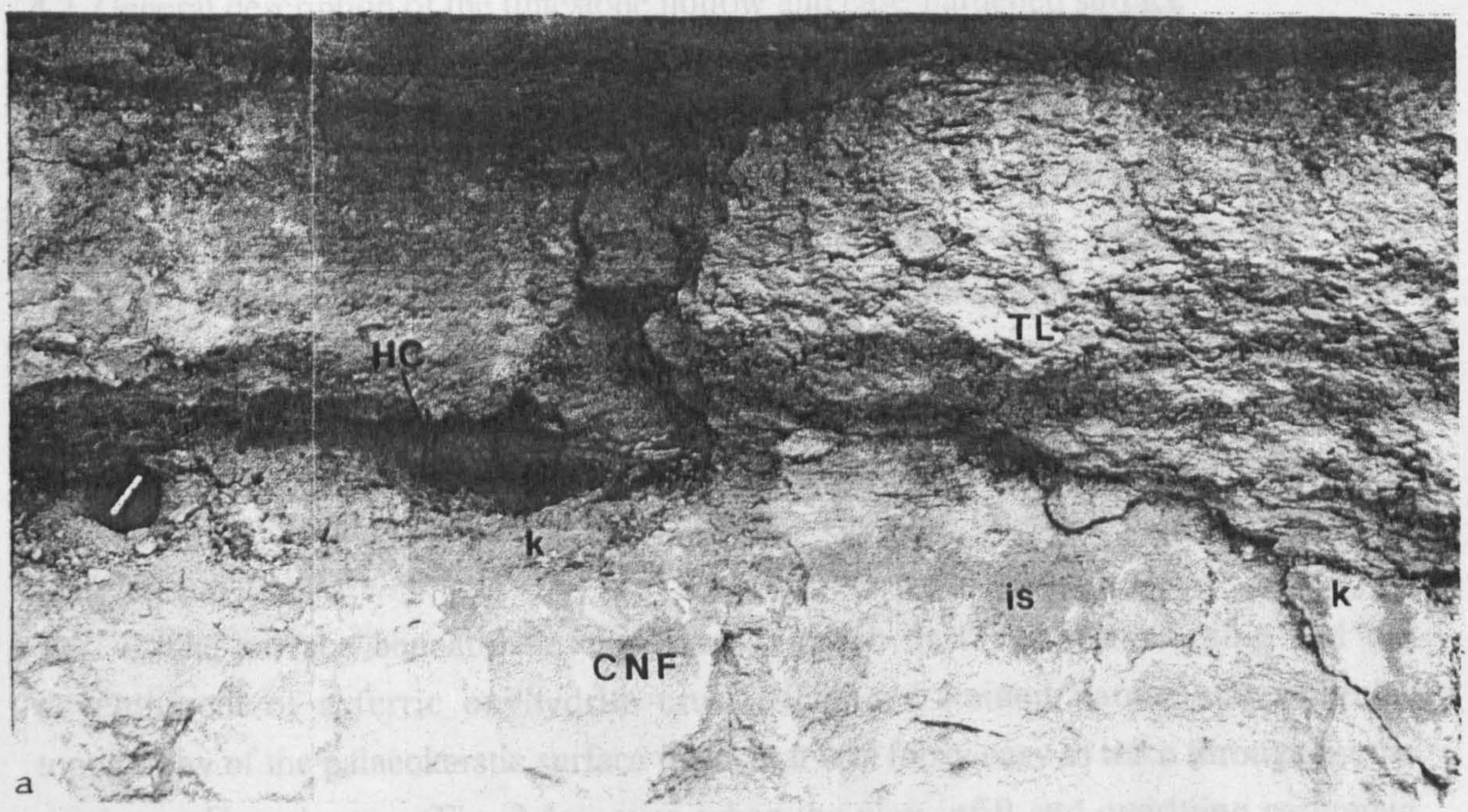
Most of the large- and small-scale features exhibited in limestone landforms are fairly cosmopolitan. However, in order to fully explain and understand the existence and formation of a karstic landform, it is necessary to study the relationship between several factors (Trudgill, 1985). These are:

(a) The original limestone rock type. For example, the compositional purity of the original limestone, and the type and degree of cementation. This is important when considering their solubility in natural waters. This is illustrated beautifully by the differences highlighted between the Jurassic oolites and Carboniferous limestones of the British Isles.

(b) The terrestrial environment and operation of erosional processes for karst formation. This includes climatic factors such as temperature and precipitation levels, as well as the amount or lack of soil cover, the pH of ground-waters and the carbon dioxide levels within ground or soil waters.

The first factor has been described in the previous chapter for the Chipping Norton Formation limestones and shall be referred to only briefly in the next sections. However, the second factor will be considered in more detail in the description of the Hornsleasow palaeokarst and infilling sediments (this and the next chapter).

Figure 4.1. (a) A small pocket of clay (HC) found some 50 metres west of the main site overlying the ferric iron-stained (i.s.) palaeokarst (k). Overlying the lens is the transgressive sand unit (TL) which is 1-1.5 metres thick and can be traced along much of the quarry section, where the clays pinch out. Other abbreviations as Fig. 2.4. (b) A field sketch of a large karstic dissolutional funnel-shaped fissure or karren within the Chipping Norton Formation limestone face some 70 metres west of the site. The karren is infilled with the soft and fissile arenaceous transgressive limestone and a few intraclast pebbles composed of underlying oolite. There is no trace of the clays in this karst feature. (c) and (d) illustrate the condition of the palaeokarstic surface underlying the clays within the hollow after excavation in Site A. The surface shows large and sharply defined cambering fractures (cf) which pervade deeply into the underlying limestone, up to several metres. The karst surface (k) also shows much less well-defined irregular dissolutional fissures, particularly noticeable in the left foreground of (c).


HC
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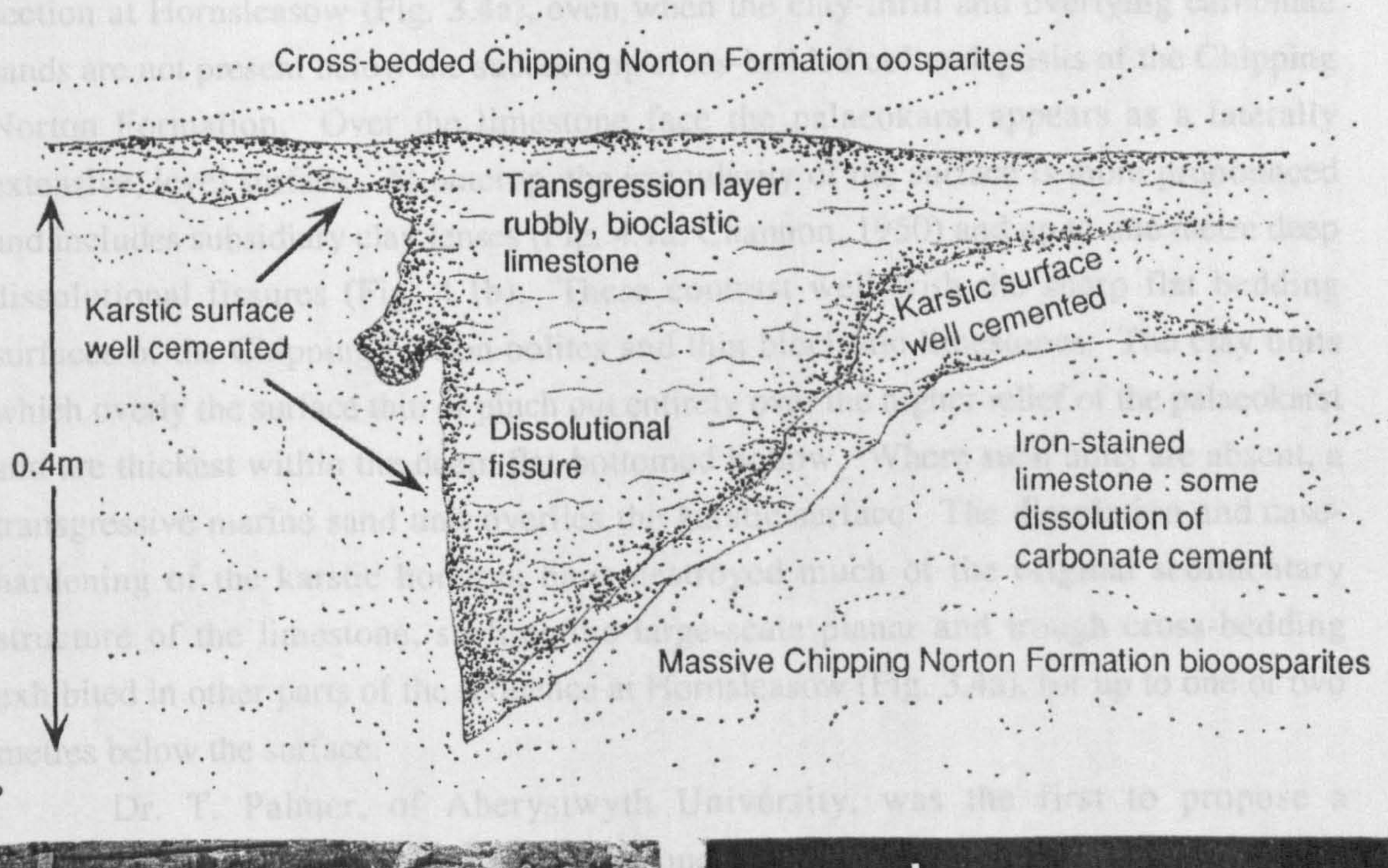
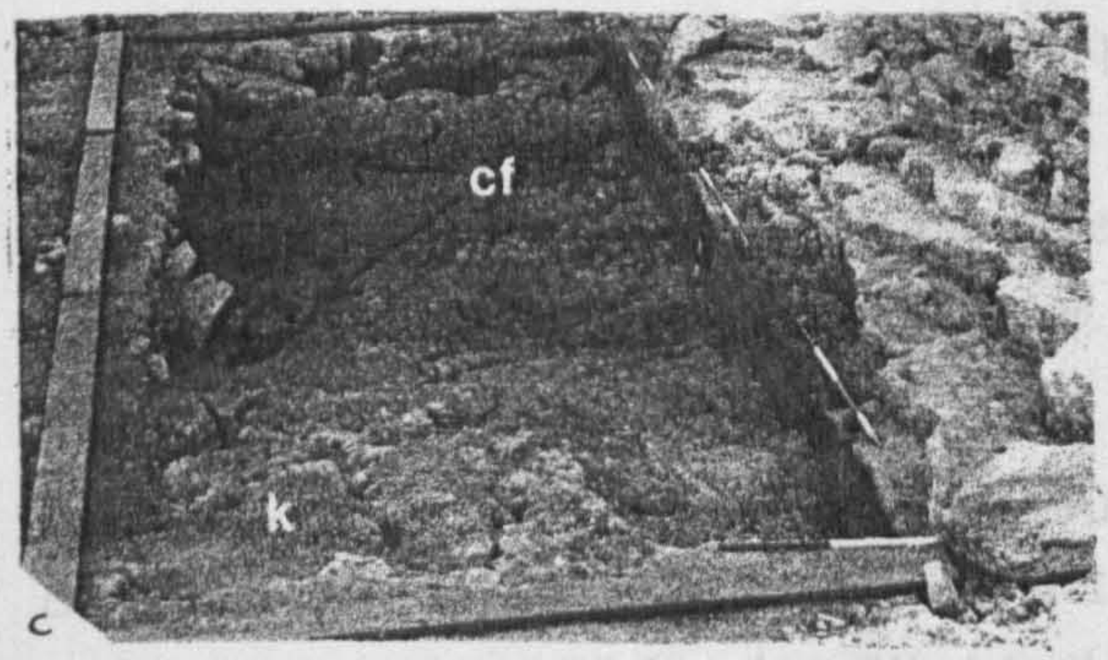
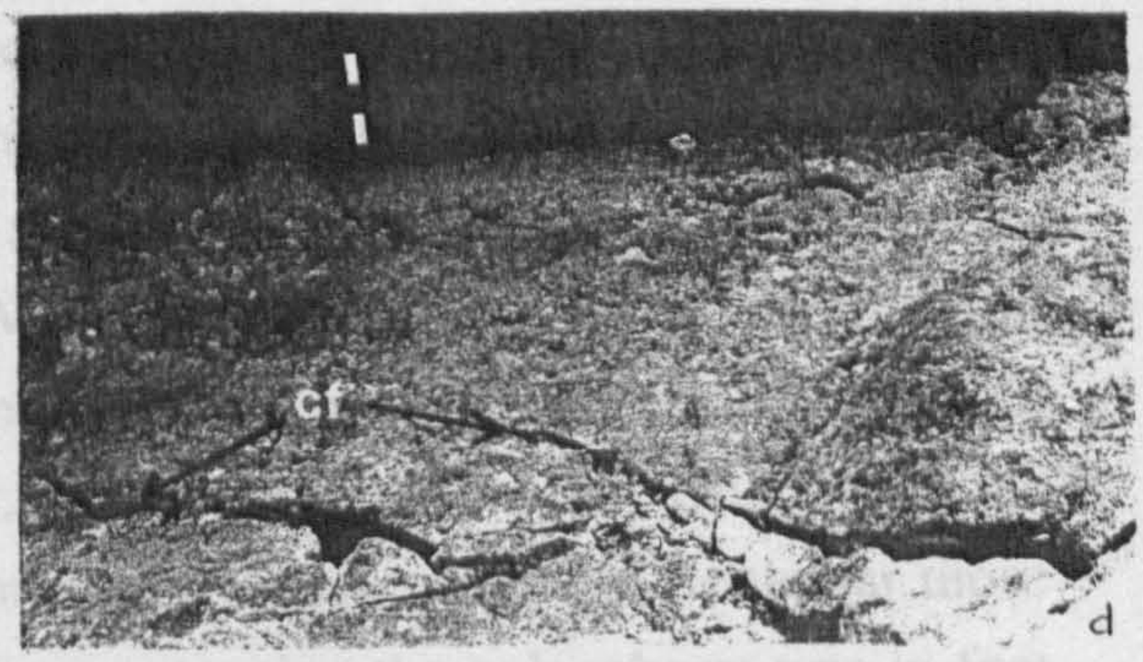

Cross-bedded Chipping Norton Formation oosparites
Transgression layer : rubbly, bioclastic limestone
Karstic surface : well cemented
Dissolutional fissure
Karstic surface well cemented
Iron-stained limestone : some dissolution of carbonate cement
0.4m
Massive Chipping Norton Formation biooosparites

b


cf
k

c


cf

d

4.2. General description of the limestone hollow and case-hardened surface

The most obvious feature of the erosional surface within the Chipping Norton limestones at Hornsleasow Quarry is the large, clay-filled, hollow from which the diverse fossil assemblage was recovered. After complete excavation of the overburden and clay, the irregular hollow was found to occupy an area of at least 16mx16m, with a depth of up to 0.7-0.8m. The hollow and irregular topography of the limestone surface is similar to features described from the undulating palaeokarstic limestone surfaces of British Dinantian (Carboniferous) sequences (Walkden, 1974, Davies, 1991), and small-scale karstic hollows produced by modern weathering of Pleistocene platform carbonates along the southwest Florida coast (Hine *et al.*, 1988). They are interpreted as original karst features.

The surface beneath the clays shows much dark red iron-staining and the development of a ferric oxyhydrite crust. The red-stained nature and irregular topography of the palaeokarstic surface mean that it is fairly easy to trace throughout the section at Hornsleasow (Fig. 3.4a), even when the clay-infill and overlying carbonate sands are not present below the succeeding cross-bedded oolite deposits of the Chipping Norton Formation. Over the limestone face the palaeokarst appears as a laterally extensive, level surface. At outcrop, the irregularity of the surface is more pronounced and includes subsidiary clay lenses (Fig. 4.1a; Channon, 1950) and up to one metre deep dissolutional fissures (Fig. 4.1b). These contrast well with the sharp flat bedding surfaces of the Chipping Norton oolites and thin bioclastic limestones. The clay units which overly the surface thin or pinch out entirely over the higher relief of the palaeokarst and are thickest within the deep, flat-bottomed hollow. Where such units are absent, a transgressive marine sand unit overlies the karstic surface. The dissolution and case-hardening of the karstic horizon, have destroyed much of the original sedimentary structure of the limestone, such as the large-scale planar and trough cross-bedding exhibited in other parts of the sequence at Hornsleasow (Fig. 3.4a), for up to one or two metres below the surface.

Dr. T. Palmer, of Aberystwyth University, was the first to propose a palaeokarstic origin for the hollow. Subsequently, Dr V.P. Wright (in Vaughan, 1989) interpreted the apparently smooth relief of the surface within the excavated hollow as evidence for development of the karst under soil cover ('covered' karst; Davies, 1991). However, more recent excavations by Dr. M.J. Simms (Simms & Metcalf, in prep., 1994; and see below) have suggested that much of the apparent smooth and rounded appearance of the surface underlying the clays, was an artefact of recent weathering of the exposed surface. The main features of the palaeokarst surface, and in particular those exposed at the base of the clay-filled hollow suggest that the palaeokarst was in fact formed with little surface soil cover (Metcalf *et al.*, 1992; i.e. 'uncovered' karst, Davies,

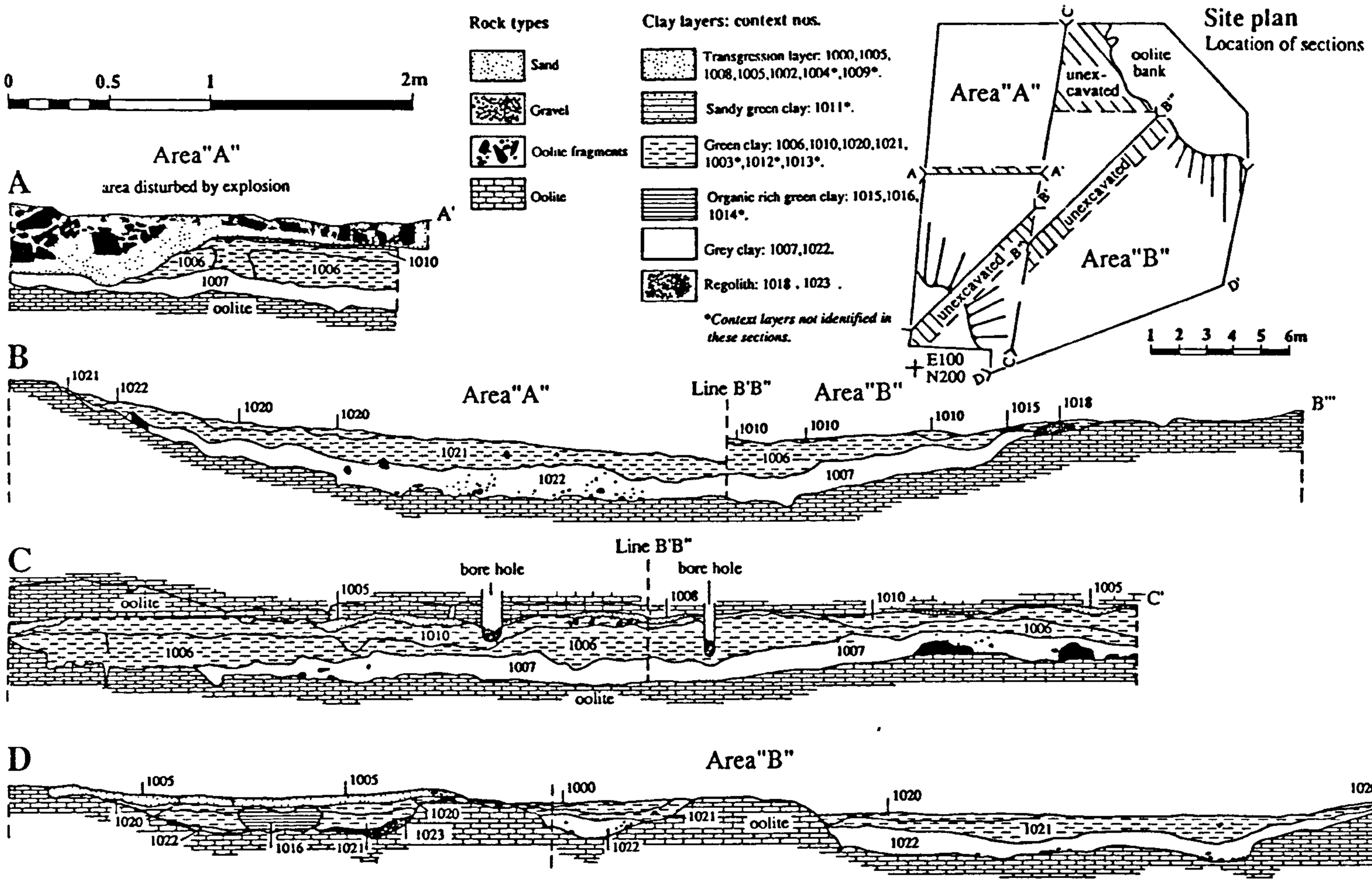


Figure 4.2. Detailed cross-sections across the clay lens based upon data collected during excavation by the archaeologists from undisturbed 'walls' of clay along various lines of intersection (shown on the site plan - inset; Fig. 2.6b). The sections show the general topography of the lens, the irregular base, and the internal stratigraphy of the clays. Context numbers correspond to different varieties of clay and limestone overburden (after Darlington, 1988).

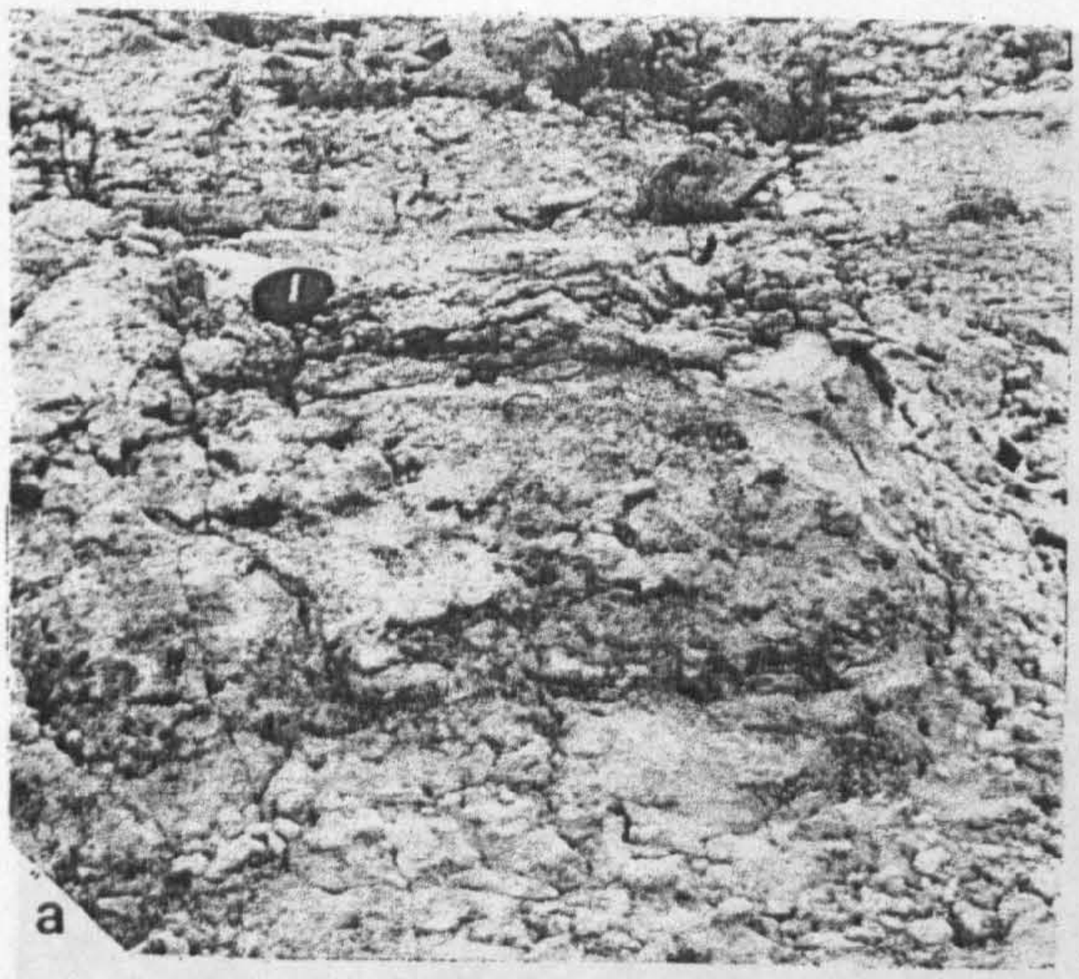
1991). Much of the descriptive and interpretative work upon the large scale features of the karst surface outlined in the following section is based upon a study made by Dr. M.J. Simms (Cheltenham and Gloucester College) in 1989. This is currently being revised in conjunction with the present authors' work upon the petrography of the limestones and overlying paleosol units and shall be referred to as Simms & Metcalf (in preparation, 1994).

4.2.1. Description of large-scale dissolutional features or karren forms

The karstic landforms and other features of the limestone are best seen in the region of the excavation site, and in particular within and around the clay-filled hollow. These were studied from photographs and site descriptions (Figs. 4.1-4.4). The shallow depression which contains the clays has an irregular outline covering about 16x16m (Fig. 2.6). In places the underlying surface slopes gently towards the centre or south east (Fig. 4.2), but at several points the sides of the hollow rise at an angle of 50-60^{o} for up to 0.5m to an outer flat rim, which is level with the surface of the upper Green Clay paleosol horizon (Fig. 4.2). Much of the floor of the hollow is irregular or gently undulose (Fig. 4.2). There are developed within the karstic surface many joints and cracks which became apparent after excavation (Fig. 4.1c, d). Although some appear to be weathered and infilled with sand, most are quite sharply defined, straight-sided 'gull fissures' and occur in all parts of the quarry where the limestone face has been allowed to degrade since quarrying activity ceased. Gull fissures are related to cambering of the limestone face within the quarry rather than being original karst features.

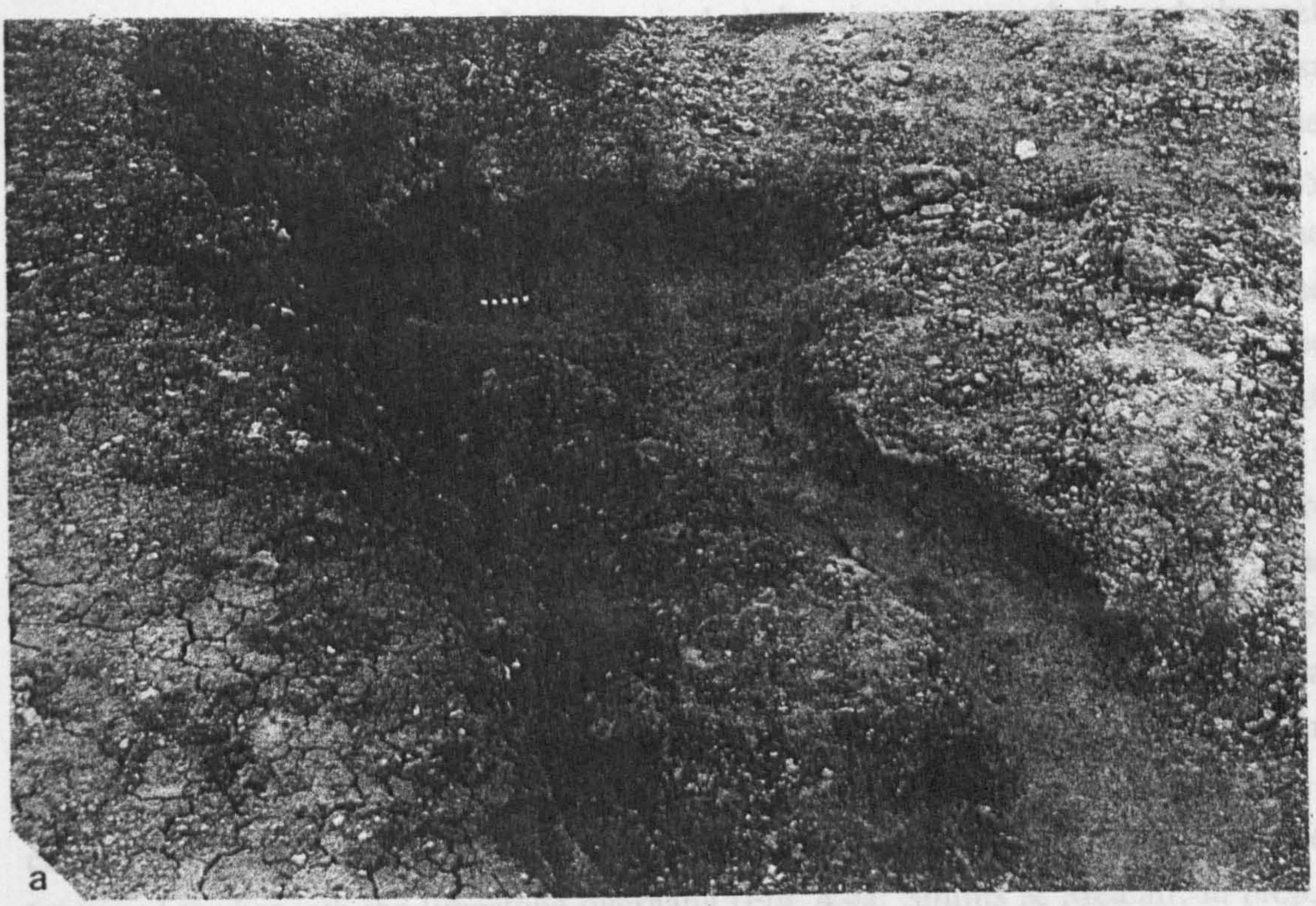
The 1988 excavations uncovered many original features of the karstic hollow, including the irregular and rubbly texture of the palaeokarstic surface underlying the clays (Figs. 4.1c, 4.3b,c). A further minor re-excavation of the slumped clays in the eastern part of the unexcavated lens, undertaken by M.J. Simms (pers. comm) in 1989, exposed an unweathered area of the underlying palaeokarst, since by then the main excavated site had degraded (for example, Fig. 4.3a taken in 1993 by the author). The site photographs shown in Fig. 4.1c illustrate the apparent smoothness of some parts of the excavated surface in site A, this is considered by M.J. Simms to be an artefact of the techniques employed during the early part of the excavation, when bone recovery was of paramount importance, rather than a true feature of the karst. Careful removal of the overlying clay from the palaeokarstic surface during excavation of Site B (Fig. 4.4) and subsequently by M.J. Simms in the same area revealed several sharp, well-preserved subaerial karstic solutional features or karren (Figs. 4.3b,c, 4.4). The sharpness of the karren forms at Hornsleasow was found to be very susceptible to mechanical erosion and similar features in the original excavation site would have been easily destroyed or obscured during the course of the excavation and by subsequent weathering.

Figure 4.3. (a) Modern weathering exposing the red iron-stained crust pervading the case-hardened palaeokarst surface at the north-western edge of the excavated site (Site B), taken in 1993. (b) knobbly case-hardened surface of the palaeokarst underlying the soil units in the excavated site A. (c) rinnenkarren and over deepened hollow developed upon the karst underneath the excavated soil at site B.


a


b


C



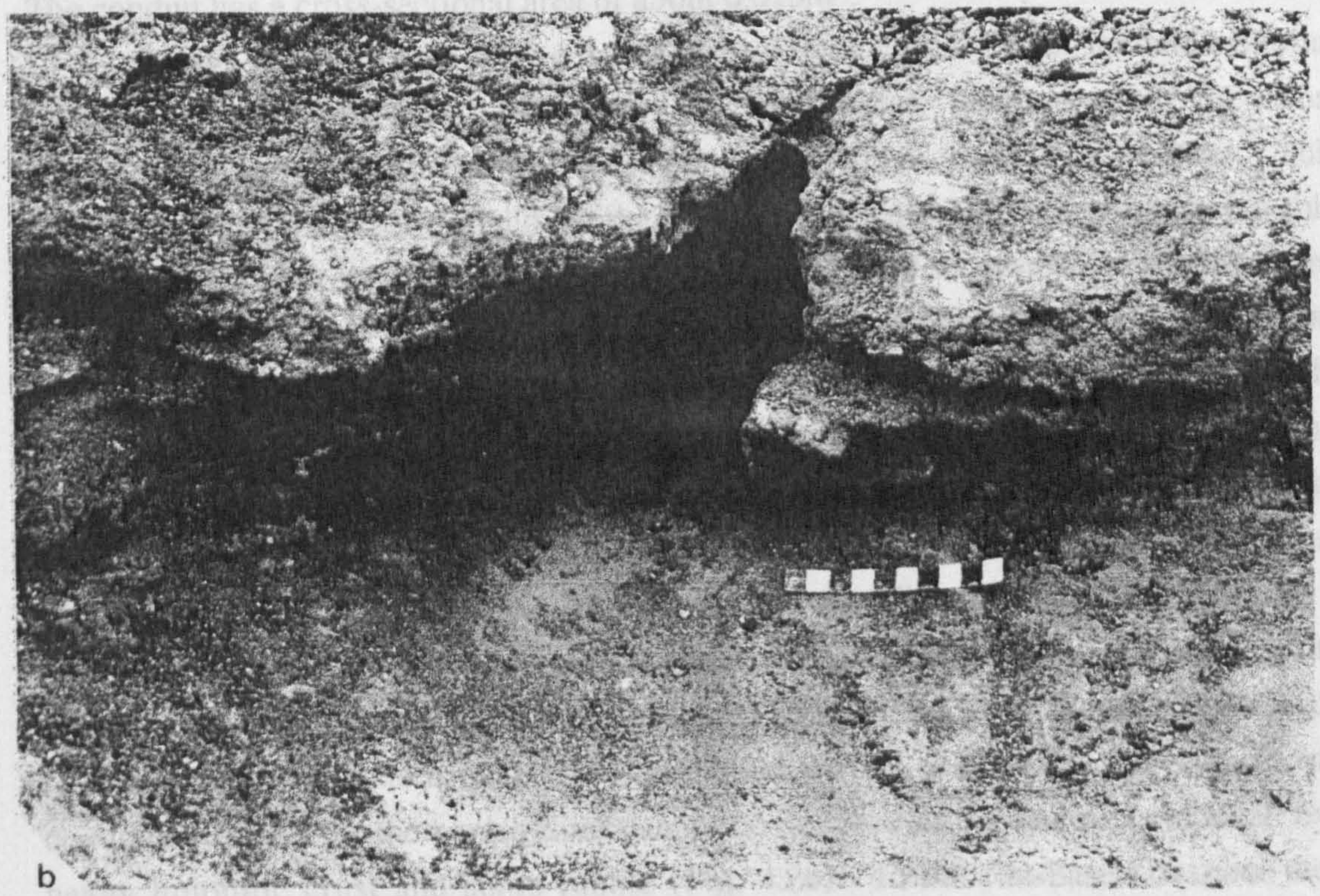


Figure 4.4. Rinnenkarren and over deepened bean-shaped hollow (a) within the palaeokarst surface exposed by a minor re-excavation of Site B, in 1989. Close-up of the case-hardened overhanging walls (b) of this hollow and crack-shaped conduit described in the text. The walls show prominent horizontal projections which are caused by preferential dissolution along bedding planes.

Karren identified by M.J. Simms included a sinuous trough-shaped channel roughly orientated northeast to southwest (Figs. 4.4a) and was discovered in the most southwestern part of the excavated Area B, against the southerly oolite walls of the hollow. The channel is about 0.15m wide and 0.8m in width, it has steep or slightly overhanging sharp-edged sides, with an angular rim (Figs. 4.4a). The floor of the channel is gently concave along most of its length, although it flattens out toward the southern end of the feature (Fig. 4.4a; Simms & Metcalf, in prep.). The depth of the channel increases from about 0.06m in the north to about 0.1m at the southern end, where it widens abruptly into an irregular, bean-shaped hollow (Figs. 4.4a).

The fabaceous hollow is flat-floored and becomes slightly overdeepened toward the southern end (Fig. 4.4b). The walls of the hollow, particularly on the southern and western edges, are steep and form prominent overhangs (Fig. 4.4b). In detail, they show substantial etching of the original limestone bedding, which forms horizontal projections (Fig. 4.4b). In the southern wall, a wide crack (0.1m in width) in the overhung wall running approximately in the same direction as the channel, reveals a conduit beneath (Fig. 4.4b). The conduit appears to run along a minor joint, but cannot be traced far beneath the limestone overhang (Fig. 4.4a; Simms & Metcalf, in prep.). The conduit has a cross-sectional area of about 0.015m^2.

Karren observed and photographed by the excavators in 1987, include a similar sinuous channel and bowl-shaped hollow (Fig. 4.3c), discovered in the most southwestern part of Site B and trending north-northeast to south-southwest. This channel is much shorter and shallower than the previous karren form and appears to split into two or possibly three subsidiary channels after 0.2m or so toward the southeast and southwest (pers. obs.). Associated with this splitting is a general shallowing of base and sides from around 0.1m in the main channel to 0.05m in the subsidiary channels. The edges are also much less pronounced in the secondary channels than in the original feature (Fig. 4.3c), which show a degree of overhang and etching along the bedding planes. The channel widens toward the northern end into an irregular bowl-shaped overdeepened hollow; with pronounced overhanging walls. The base of the hollow is obscured by rubble and heavy iron staining (Fig. 4.3c), but is concave, and slopes gently toward the back northern wall. There appears to be the remains of a small (10mm wide) fissure in the northeastern face. However, whether this crack holds the remains of a conduit and drain for the karren form cannot be deduced as the feature is blocked by a heavy limonitic crust and some rubble (Fig. 4.3c). In the flat-based channel there appears to be the remains of a small runnel in the southeastern corner of the hollow which extends southwards along the channel floor (Figs. 4.3c).

4.2.2. Interpretation of large-scale karstic features

The clay-filled hollow at Hornsleasow is interpreted as a solution doline developed upon a low hummocky karstic horizon. Similar 'medium-scale' (i.e. metre and centimetre scale) topographic features have been noticed in the coastal karsts forming today along the Florida coastline in the Gulf of Mexico (Hine *et al.*, 1988). These sinkholes are usually no more than a metre in depth and are commonly developed into freshwater ponds in the coastal swamps (Hine *et al.*, 1988, p. 569). They are formed by surface dissolution of the limestones. The coastal karst of northwest Florida also possesses much larger scale karstification features, such as tidal creeks, large hummocks which form islands in the swamp or 'marsh hammocks' and even the large embayments at the mouths of rivers draining into the Gulf (Hine *et al.*, 1988). The limited extent of the palaeokarst at Hornsleasow has meant that large scale features and subsurface karstification (such as cave passages) have not observed.

The features of the Hornsleasow palaeokarst are interesting because they seem to be closely analogous to karstic forms developing today upon areas of well-cemented and impermeable limestone with little soil cover. This is in direct contrast to the karstic features developed at the present within the weakly indurated and porous oolites of the Middle Jurassic in the Cotswolds. For example, the steep side-walls of the clay-filled hollow and those of subsidiary dissolutional fissures (Fig. 4.1b), the well-indurated horizontal bedding projections and the conduit, are quite unlike the rather rounded and ill-defined karst forms of this area and others karstic sequences overlain by fossil soils (e.g. Harrison, 1977; Simms & Metcalf, in prep.). The channel features within the base of the hollow are the most interesting karren form, as morphologically these resemble 'rinnenkarren', or solution runnels which develop as small stream channels on bare rock slopes (Ford & Williams, 1989).

Similar karren forms have been observed by M.J. Simms in the Burren of County Clare, Eire. He describes well-developed rinnenkarren, virtually identical to that described above, which occur on the extensive tract of limestone pavement in the broad valley situated to the north of the Carha Bridge, on the Fanore-Lisdoonvarna road, in the Caher River valley (Simms & Metcalf, in prep.). However, the rinnenkarren seen in the Burren are developed from temperate karstification of well-indurated ancient shelf bioclastic limestones, which have very low primary permeablility and low associated porosity, but significant secondary permeability along joints and bedding planes. Hence, any solution of the limestone must occur along joints and stress fractures, so developing into the characteristic 'clint and grike' limestone pavement scenery typical of the Burren (Trudgill, 1985). The similarity of the rinnenkarren developed on the Burren with those seen at Hornsleasow, might be suggestive that the limestone bed on which the Bathonian

karst was developed, was significantly more indurated and of lower permeability than the other limestones in this area.

The petrographic sections (see below and section 3.5) of the karst limestone and those of the surrounding Chipping Norton Formation limestones do not indicate that there is any significant difference in composition or lithification for this bed and if anything, at present much of the bed directly underneath the clay-filled hollow is de-calcified and poorly cemented at the surface. However, the induration associated with parts of the rubbly surface developed around the karren forms could be the remains of beachrock formation or a 'case-hardening' crust associated with the early sub-aerial development of the palaeokarst and analogous to the induration which occurs in some freshly exposed modern carbonates (cf. Bathurst, 1975). Trudgill (1985) and Ford & Williams (1989) have described how sub-aerial exposure of young, weakly indurated limestones in warm temperate or tropical climates often leads to reprecipitation of calcium carbonate leached from the limestone as a hard aragonitic or calcite crust upon the surface of the porous limestone. Beachrock lithification takes place beneath the ground surface, although it can be exposed through storm erosion and forms in the intertidal or supratidal conditions (Fig. 3.7) and is initiated by evaporation and CO_2-degassing of seawater which is flushed through the sediments in the marine vadose zone (Tucker and Wright, 1990). Either phenomena imparts a much harder and considerably less permeable texture to the limestone bed in a zone extending for around one to two metres below the surface. It is also possible that the palaeokarst horizon may have originally been cemented as a hardground in the subtidal environment (section 3.6.1). Whatever its origin, the early lithification of the Hornsleasow surface would then have allowed the development of karren forms during karstification which would not have formed upon the otherwise weakly cemented and relatively porous marine limestones.

Rinnenkarren typically are developed on bare, gently inclined limestone surfaces. The example described by M.J Simms at Hornsleasow (Fig. 4.4), lies at a topographical low point on the palaeokarst surface and slopes to the southeast before broadening into a shallow, overdeepened depression or sink, flanked by the low steep-edged walls (Simms & Metcalf, in prep.). The solutional etching along the bedding which is developed around the depression may have occurred at times of high runoff, when the conduit at the base of the sink was unable to drain the area at a sufficient rate to prevent ponding. The conduit and rinnenkarren itself appear to follow a minor joint, indicating that the limestone surface was already well-lithified with some fracturing developed prior to the main period of karstification. It seems probable that the channel and associated conduit formed a drain to the karstic hollow (see section 4.7).

The main bowl-shaped clay-filled hollow and the other subsidiary hollows can be described as discrete closed depressions, rather than dry river channels. Closed depressions within karstic regions are known as 'dolines' (Trudgill, 1985) and there are

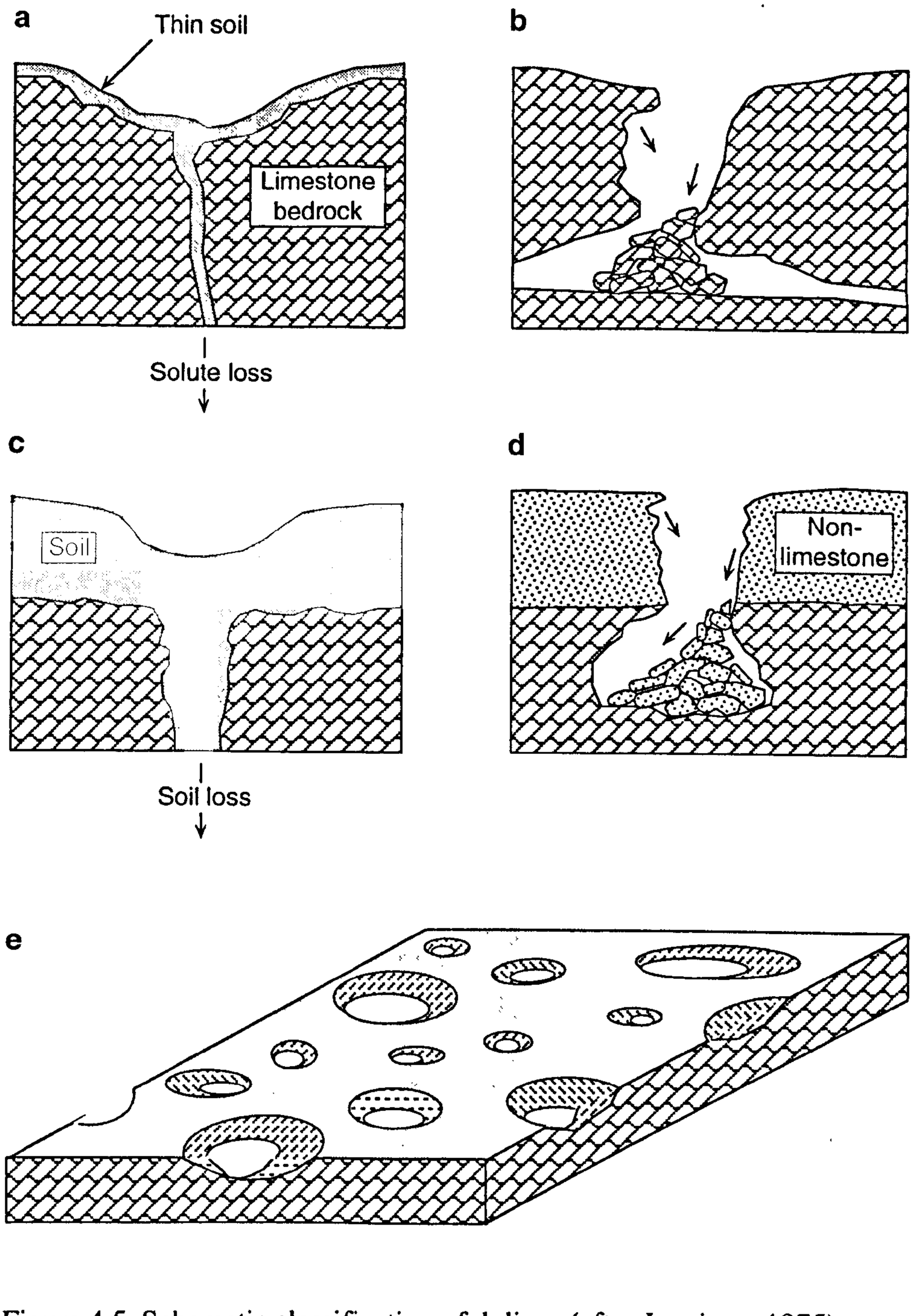


Figure 4.5. Schematic classification of dolines (after Jennings, 1975), showing: (a) solution doline; (b) collapse doline; (c) alluvial doline; (d) subjacent karst collapse doline. See text for explanation. (e) A schematic cross-section of a doline karst landform (after Jennings, 1985).

four major types of doline, which originate from different processes (Sweeting, 1972, Jennings, 1975; Fig. 4.5), these are: (1) 'solution dolines', where bedrock is dissolved and lowered, with solute loss down a conduit; (2) 'collapse dolines', where there is collapse and physical movement of limestone bedrock into solution cavities or cave passages; (3) 'alluvial, drift or subsidence dolines', were solutional of the bedrock takes place underneath a substantial soil cover, with associated soil subsidence into the hollow and loss down bedrock cavities; (4) 'subjacent karst collapse dolines', where non-limestone rock forms a roof above limestone and collapses into solutionally opened cavities within the limestone bedrock. In the absence of any evidence for large-scale cave passages below the Hornsleasow karst, the hollow would appear to be a solutional doline (Simms & Metcalf, in prep.).

4.2.3. Description and interpretation of other features associated with the karst

The palaeokarstic surface as seen within the limestone faces at Hornsleasow is readily distinguished from other horizontal features such as bedding planes, by its irregular hummocky nature (Fig. 4.1a) and the strong reddened appearance of the rock up to 0.2-0.5 metres below the surface (Figs. 4.1a, 4.3a). The red-staining of the rock appears to be pervasive (Fig. 4.3a) and is associated with a ferric oxyhydrite or goethite crust, and importantly in some places, a thin haematite crust (section 4.4). The floor (Fig. 4.3c) and rim (Fig. 4.1a) of the excavated hollow is in many places coated with a hard yellowish-brown limonitic or reddish-brown haematitic crust. However, somewhat enigmatically, the steeply inclined sidewalls of the hollow appear to lack any evidence of this crust (Simms & Metcalf, in prep.).

The limestone floor of the hollow is fairly friable and soft (Fig. 4.4), and has degraded much since the initial excavation of the overlying clays. This is illustrated by the rubbly but clearly defined nature of the karren forms (Fig. 4.3c, 4.4a,b) and surface of the karst (Fig. 4.3b), which were photographed during the original excavations (1987-1988), and the exfoliation of the ferric oxyhydrite crust and underlying limestone at the edges of the hollow, exhibited in the photograph which I took some five years later in 1993 (Fig. 4.3a). Although much of the surface surrounding these features is now only weakly cemented, the sharpness of the karren forms (Figs. 4.3c, 4.4a) and associated bedding projections (Fig. 4.4b) suggests that the original surface was well-indurated. Where the surface lacks an overlying clay cover elsewhere in the quarry, the palaeokarst forms a prominent feature and is much better cemented than the overlying transgressive sand unit (Fig. 4.1b).

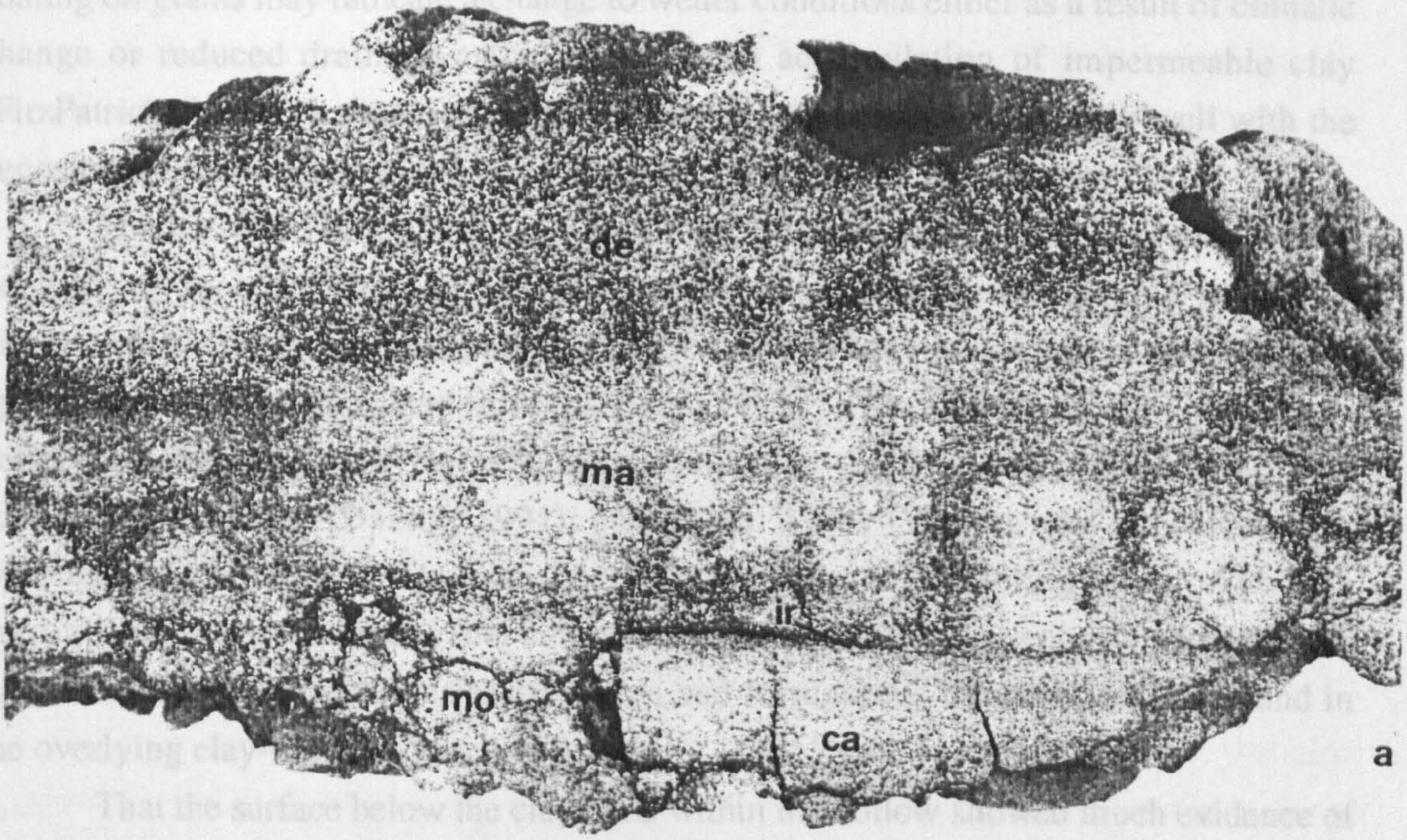
The indurating cement is preserved in some places within the soil-filled hollow. For instance, the knobbly appearance of the floor of the hollow (Fig. 4.3b,c) could be a function of this early cementation and the subsequent preferential de-calcification. This is indicated by the fact that many pebble-sized clasts of well-cemented limestone were

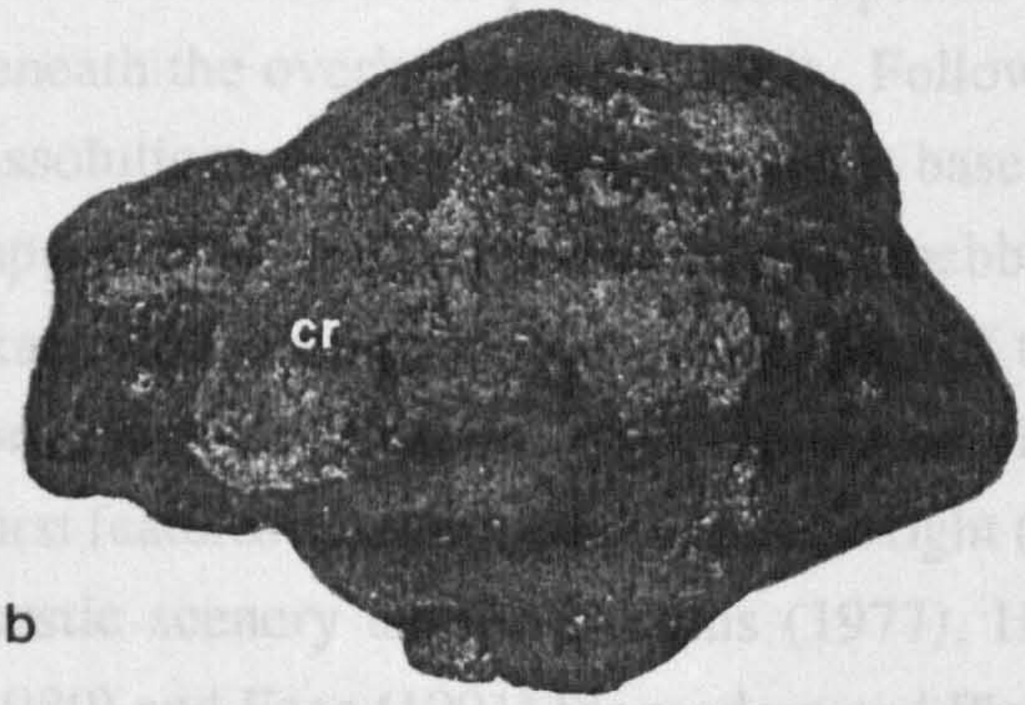
discovered lying upon the karst surface during the excavation (Fig. 4.6b-e). These clasts are on the whole of similar composition to the underlying rock type (i.e. intraclasts: see section 4.4). They show an dark iron-stained indurated outer "skin" and some rounded examples also possess a well-developed polish. Most of the clasts appear to be intraclast pebbles formed in conjunction with the indurated palaeokarst, although some are pedogenic nodules from the overlying paleosols. Other clasts exhibit less abraded broken surfaces, which may indicate that they were once attached as indurated 'knobs' to the surface of the karst. Intraclasts are common in both marine hardground, intertidal beachrock and case-hardened crust horizons, and form through brecciation and erosion of the indurated limestones (Tucker and Wright, 1990). Similar darkened clasts known as 'black pebbles' have been described from both modern (e.g. Strasser, 1984; Shinn & Lidz, 1988) and ancient exposed karstic sequences (e.g. Strasser & Davaud, 1983).

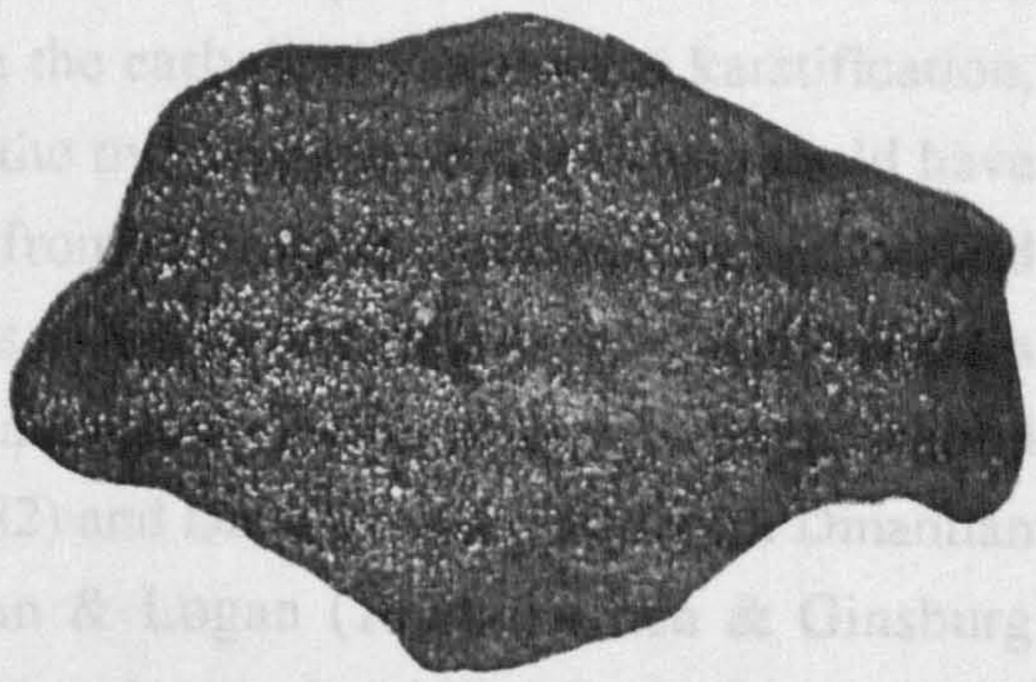
Most of the limestone pebbles studied by the author were collected by M.J. Simms from the unexcavated eastern part of the hollow, which suggests that they had become separated from the underlying rock prior to excavation disturbance. Although most of these examples were derived from the weathering of the underlying limestones, there are one or two (Fig. 4.6e) which are composed of allochthonous limestone material. The exotic clasts are similarly well-rounded and exhibit the same iron-stained polished surface as those derived from the underlying karst. The presence of exotic clasts and the well-rounded nature of the pebbles suggests that there were frequent flooding events unloading transported sediment into the hollow (section 5.1). The best preserved indurated karstic surfaces are found beneath some of the large cetiosaur bones, which are themselves cemented to the underlying surface by the calcitic cement (section 10.5.1). The bones are also pervaded by fine calcite veins and exposed cancellous tissue is infilled with this cement.

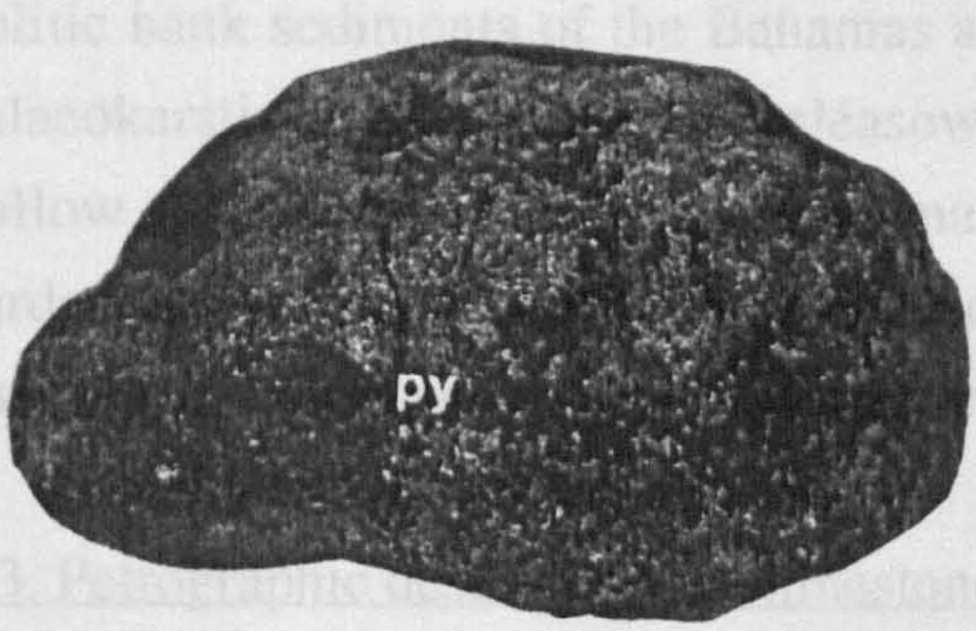
The majority of the limestone clasts found upon the karstic surface of the hollow appear to be well-rounded. Their smoothed polish was probably initiated by subaerial weathering, as well as mechanical rounding caused by rolling around in the hollow prior to stabilisation by deposition of the clays. It is interesting to note here that most of the *Cetiosaurus* bones also show a high degree of *in situ* weathering before soil deposition and can best be described as 'bone pebbles' (section 10.4.2).

A fine-dusting of the black dendritic manganese oxide (MnO_2) mineral 'pyrolusite', is seen upon the better-preserved surfaces of the karst. This dusting is easily removed by abrasion or subaerial exposure, and hence, was effectively removed from much of the karstic surface during and after the excavations. It is preserved upon the upper surfaces of the cetiosaur bones and also those of the limestone pebbles (Fig. 4.6d). Manganese dioxide is a common component of lateritic weathering, found in association with iron oxides and other weathering or detrital products (Foos, 1991) and is a feature of clasts occurring within moist or seepage conditions. Patches of pyrolusite









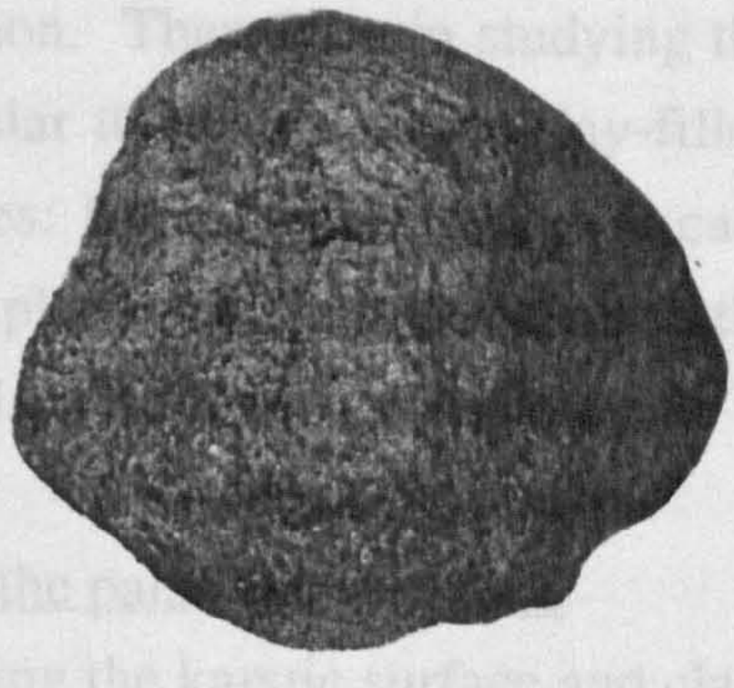


Figure 4.6. Polished section through top 150mm of palaeokarstic horizon (a) on HQCN3, showing the regions of different alteration (see Fig. 4.7 and text). Karstic pebbles (b-e) taken from the base of the clay-filled hollow. (b-c) HQP3, length = 55mm; (d) HQP4, length = 45mm; (e) HQP6, length = 30mm. Abbreviations: 'ca' case-hardened limestone; 'cr' ironstone crust; 'de' decalcified surface zone; 'ir' ironstone veins; 'ma' marbled zone; 'mo' mottled zone; 'py' pyrolusite mineralisation.

coating on grains may indicate a change to wetter conditions either as a result of climatic change or reduced drainage associated with an accumulation of impermeable clay (FitzPatrick, 1993). It would seem that the latter hypothesis ties in rather well with the depositional environment of the waterlogged Grey Clay paleosol (Chapter 5).

The karstified limestones show no strong evidence in field or thin sections of the presence of root structures or terrestrial burrowing, although the general pattern of dissolution channels and brecciation fractures seen below the surface may have been opened or utilised by a rooted flora (cf. Foos, 1991). This is in contrast to other well-defined modern and fossil karst sequences, such as the Dinantian coastal limestone scenery of Anglesey (Davies, 1991), and South Wales (Wright, 1987), which show well-developed penetration by plant root systems and a soil-dwelling fauna. Although burrows, roots and calcareous root concretions or 'rhizocretions' are not found within the karstified limestones at Hornsleasow, roots and ferruginous rhizocretions are found in the overlying clay-rich soils (section 5.4.3).

That the surface below the clay units within the hollow showed much evidence of dissolution and subsequent erosion, points to further development of the karstic horizon beneath the overlying clay deposit. Following the early lithification and karstification, dissolution must have occurred at the base of the overlying paleosol. This would have separated the limestone nodules and pebbles from the underlying indurated karst and exaggerated the overall bumpy nature of the surface in the hollow and in the process formed an overlying 'regolith' of mixed clay and carbonate detritus. Similar covered karst features have been noticed by Wright (1982) and Davies (1991) in British Dinantian karstic scenery and by Perkins (1977), Hagan & Logan (1974), Beach & Ginsburg (1980) and Foos (1991) for modern and Pleistocene karsts developed in the lagoonal and oolitic bank sediments of the Bahamas and Florida region. Therefore, in studying the palaeokarstic limestones at Hornsleasow and in particular those within the clay-filled hollow, one has to consider two formational processes: (1) initial emergence, case hardening and dissolution of an uncovered karst; (2) complex pedogenic alteration of the underlying limestones through an overlying clay paleosol, i.e. covered karst.

4.3. Petrographic descriptions of limestones surrounding the palaeokarst surface

In order to ascertain alteration of the limestones underlying the karstic surface and clay-rich paleosols at Hornsleasow, a petrographic study was undertaken. In all, five hand specimens of the Chipping Norton limestone surrounding the hollow were sampled by the author in 1990 and M.J. Simms in 1989. They were removed from the site in the standard means and their relative positions are marked upon Fig. C1 (Appendix C1). The specimens taken from the palaeokarst surface include HQCN2, a sample of limestone which I removed from the base of the excavated karstic hollow (Site B) in 1990, HQCN3, removed by M.J. Simms from the base of the hollow during his minor re-

excavation of the eastern part of Site B in 1989, and HQCN4, a sample of limestone which I took from the northern side-wall of the hollow in 1990. As with the non-karstic limestones, these specimens were prepared as polished sections, stained acetate peels and thin sections. Two sections were prepared from the top and base respectively of HQCN2 and HQCN3. Also sectioned were two calcareous clasts (HQP1 and HQP2), collected from the base of the hollow by M.J. Simms in 1989.

The full petrographic descriptions are given in Appendix C6. The primary concern of this examination was to attempt to identify and differentiate between features in the limestone which were primary, those which had occurred during pre-karst lithification and diagenesis, those that had been initiated by uncovered and soil-covered karstic weathering and finally, features arising from post-karst alteration.

4.3.1 Description of hand specimens

As with the unaltered Chipping Norton Formation limestones (section 3.5), the hand specimens, polished blocks and stained acetate peels were examined first in order to facilitate recognition of large-scale textural and mineralogical features. These included relict lamination or bedding, and an overall approximation of the limestone grainsize distribution and preliminary recognition of any large clasts or allochems (Tucker, 1981). Any jointing and mineral veins present were also noted at this point. General alteration features of the limestones examined included an assessment of the degree of cementation and weathering of the rock. This incorporated an appraisal of the colour variation across the polished transverse sections and correlation with mineralogy as determined by examination of the stained acetate peels of the same area.

All of the specimens, except one of the karstic clasts HQP2, are equigranular calcarenites typically composed of ooliths, peloids and shell fragments and are considered to have been derived from the same limestone horizon. They are typically similar in composition to the calcarenites of the unaltered Chipping Norton Formation below the karst described in sections 3.5 and 3.6.1. The clast, HQP2, is from an oolitic marl and can be considered to be a calcilutite (Tucker, 1981).

However, almost all of the specimens show varying degrees of decalcification and are in varying states of cementation. Decalcification of the cements and grains in the rock leads to the opening of solutional pathways, and the liberation of residual weathering products (e.g. quartz, organics, aluminosilicates and iron oxyhydrites: Esteban and Klappa, 1983; Foos, 1991). Such products and also detrital particles deposited on the limestone surface may become incorporated in the decalcified horizon at the surface and this process is visible upon the hand specimens in terms of colour variation and zoning, and the friable nature of the decalcified regions of the rock (Figs. 4.6a & 4.7). The karst specimens were taken from the top 0.3m of the bed which has undergone karstification, and therefore, they show segregated zones of chemical

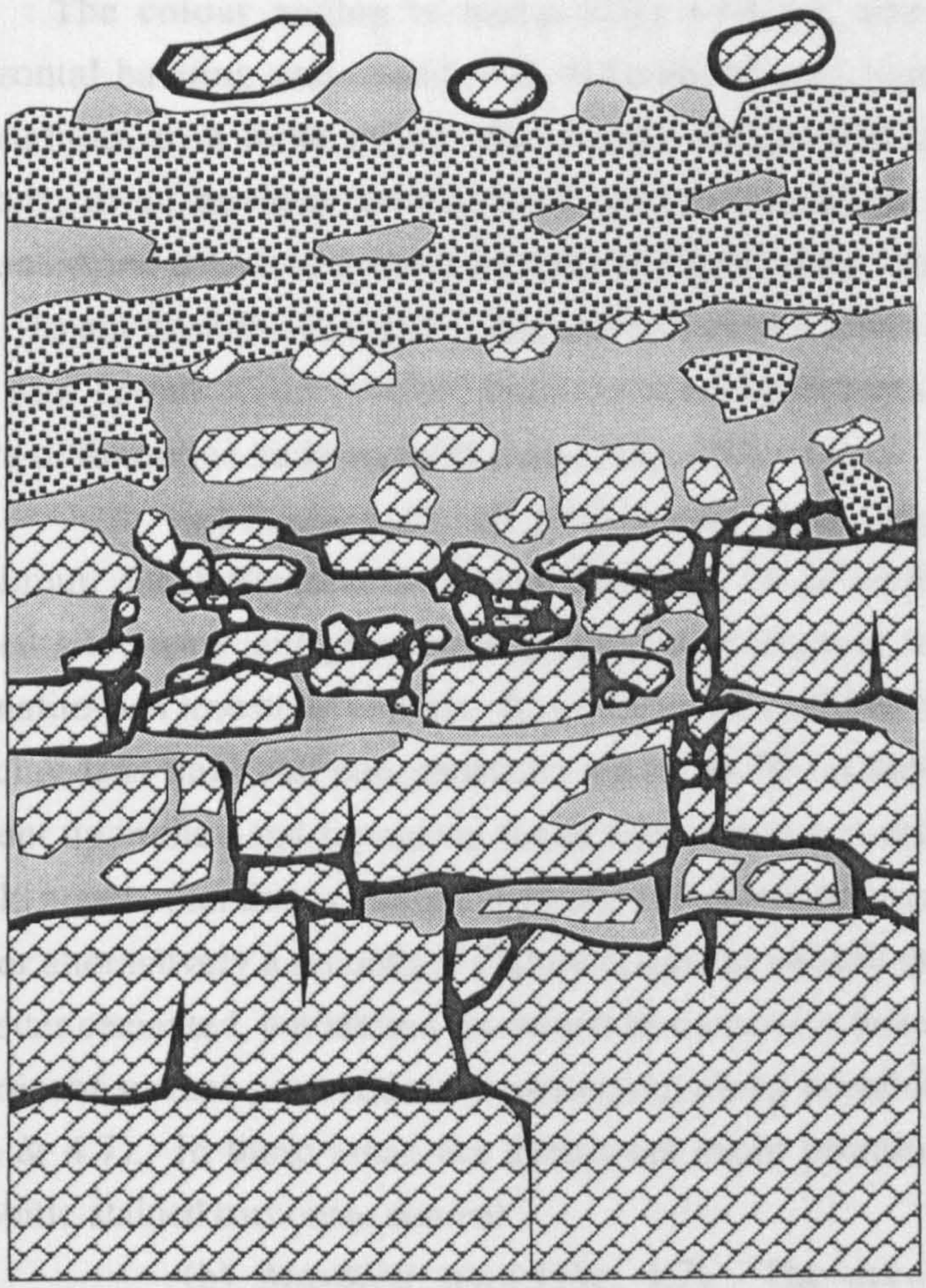


Figure 4.7. Schematic cross-section through uppermost one metre of the palaeokarstic surface underlying the soils at Hornsleasow, based mainly upon the polished sections, HQCN2 and HQCN3 (Fig. 4.6a).

alteration from the surface downwards (Figs. 4.6a & 4.7). For instance, the two specimens taken from the floor of the hollow, HQCN2 and HQCN3, are distinctly zoned and this zoning is picked out on the polished surface of the cut transverse sections by iron-staining colour variation, and to some degree, by the secondary colour variation observed on the stained acetate peels (Figs. 4.6a & 4.7).

The colour zoning is particularly striking, and it ranges from an obvious horizontal banding associated with differential cementation patterns along beds and laminations, to a more subtle and diffuse marble-like zonation in poorly-cemented regions, to an obvious 'mottling' of colour associated with areas which are better cemented and therefore decalcified preferentially along cracks and bedding planes (Fig. 4.7). To aid identification and elucidation of the processes involved, the zoning was split into four colour-zones (shown upon the schematic cross-section through the top one metre of limestone underlying the clays, Fig. 4.7):

(a) *Surface decalcified friable zone* (Fig. 4.7). This is the zone of maximum decalcification and decementation. It is typified by remnant deep-yellow etched allochems and prominent quartz grains loosely cemented by a reddish-brown haematite-rich ironstone cement. It occurs in the surface region of the blocks underlying the clay lens only and can penetrate up to 50-70 millimetres into the rock. It is not present upon the karst limestone taken from the outer wall of the hollow (HQCN4). It would appear that this is either a primary karstic phenomenon protected by the overlying soil or alternatively a secondary feature related to subsoil erosion. Remnant diffuse areas of better cemented, but altered limestone also occur in this region and these may occur as prominent surface projections, or strung out along possible ghost bedding-planes (Figs. 4.6a & 4.7). In these areas the grains are more prominent and are surrounded by a limonitic-stained carbonate cement.

(b) *Marbled zone* (Fig. 4.7). The decalcification and iron-staining reduces further into the section and gradually gives way to a zone of more diffuse alteration about 50-150mm into the rock. This is made up of areas of decalcified haematitic-ironstone, yellow-brown limonite-stained altered limestone and smaller regions (typically 10mm^2) of creamy-white 'well-cemented' limestone (Figs. 4.6a & 4.7). The boundaries of these areas are quite diffuse, giving the rock a 'marble-like' appearance.

(c) *Mottled zone* (Fig. 4.7). At about 100-150mm into the rock underlying the soil, and closer to the surface in other parts of the palaeokarst, the rock is characterised by well-defined regions of unaltered, well-cemented limestone surrounded by narrow 'veins' of limonite-rich ironstone (Figs. 4.6a & 4.7). The veins typically infilled with a dark brown material , which has hardened into a fairly good cement closing the veins off to further dissolution. They are about 1-2mm in width. The whole region has the appearance of a well lithified soil or saprolitic horizon, with the limestone

'peds' surrounded by the narrow zones of decalcification. Toward the top of the region the limestone 'peds' are fairly small and rounded and the surrounding veins are pervasive over the whole region (Fig. 4.7). However, further down the rock the decalcification becomes much more selective and the veining appears to occur along joint and bedding planes, here the 'peds' of limestone are much larger and regular in appearance (Fig. 4.7). This region can be described as a zone of moderate alteration and brecciation with well-defined veins or sutures occurring within largely well-cemented limestone. A similar feature occurs in the platy or nodular weathering zone of modern karstified limestones described by Hine *et al.* (1988) underlying coastal marshes in northwest Florida.

(d) *Unweathered limestone* (Fig. 4.7). This is the well-cemented sparry calcarenite. In the areas underlying the soil, the unaltered limestone occurs about 0.7-1.0m below the surface, however in HQCN4 and also within the pebbles, this 'unaltered' limestone occurs almost directly below a thin-skin of decalcified ferric oxyhydrite ironstone.

The above description and schematic representation of the palaeokarstic surface cannot be considered to be definitive. There is much variation on this theme throughout the sections, with some zones being absent or reduced, whilst others are much more pronounced. The scheme does seem to hold true for most of the limestone underlying the soil units at Hornsleasow. However, the small pebbles which were also found beneath the soils do not show a deep zone of decalcified haematitic ironstone, but possess instead an indurated limonitic crust. This crust is only a few millimetres thick and appears to be similar to the veins described above in being made up of ferric oxyhydrites and limonite. In some places, the crust has been removed and the underlying limestone is etched and quite friable, with grains quite prominent upon the surface. The interior of the pebbles seem to be much less altered and quite well-cemented. Variations in cementation also occur along the main paleokarstic surface, as a piece of limestone taken from the sidewall of the karstic hollow (HQCN4), is also well-cemented, and although decalcification does occur at the surface, the rock does not have an iron-stained weathering rind developed.

4.3.2. Thin section description procedure

The thin sections were examined in daylight and under the microscope, in conjunction with stained acetate peels. As with the unaltered limestones described in section 3.5 the daylight study was to recognise large-scale features, such as bedding and general grainsize distribution. This also allowed recognition of the zones of different cementation (although it must be noted that colour variations in thin section appear much paler than in the polished sections). The thin sections were prepared to differentiate between the regions of different cementation and alteration. So, for example, the specimen HQCN3 had two sections cut: one from the upper decalcified zone and the other from the lower mottled well-cemented zone. In these sections different areas of cementation were

searched for and described separately, hence for section HQCN3-1 there are two descriptions: one of the pale-yellow (in thin section) poorly cemented opaque-rich limestone and the second, of the mostly decalcified haematitic ironstone regions.

The regions of the limestone which have undergone alteration during the period of karstification and pedogenesis exhibit a mixture of features inherited from the parent rock, such as composition of grains and cement, and others formed during this period, such as weathering products and new cements. The descriptions of the 'unaltered' limestone areas provide a basis on which to form a theory of how and in what way has the karstic and pedogenic weathering altered the original rock composition and texture. In attempting a simple classification scheme of the limestones, the most applicable characters from both the descriptions of limestones and soils were used (Retallack's scheme outlined in Chapter 5 and Appendix C5). The description of the thin sections proceeded along the following guidelines:

(1) Classification of original features. These include the inherited texture and composition of the rock, which are based upon the Folk and Dunham schemes for limestones (section 3.5.1). This aims to illuminate any difference between the altered and parental limestones. As with the other limestone beds at Hornsleasow, cementation history, porosity and other processes such as neomorphism, and pressure-solution are also described.

(2) Classification of secondary features. The preservation of grains was examined and in some areas it provided a rough appraisal of the degree of alteration of grains, although in the more altered zones this was to prove a futile task. Alteration products of carbonate grains, such as micritised allochems and ferric oxyhydrite residues, were looked for and the degree of alteration and weathering of allochems was also ascertained. Finally, the cementation history of the altered region was ascertained, in terms of relict primary cementation, karstic and pedogenic alteration of primary cements and post-karst cements. At this point the porosity of the rock was also examined, as this can form from secondary dissolution as a result of diagenesis or sub-aerial karstic and pedogenic weathering, as well as being a primary depositional feature of the limestone (section 3.5.3). Finally, the structure or total organisation and spatial distribution of the most altered zones in thin section was also described. Structure is a term used in soil micromorphology and is associated with the porosity of soils (FitzPatrick, 1993). However, it can be applied to any disaggregated rock type and was particularly useful in describing areas which were almost completely decalcified.

This procedure was followed for each of the zones of alteration described in the study of the hand specimen (sections 4.3.3-4.3.7). The interpretation of these zones as a palaeokarstic weathering profile is given in section 4.3.8.

4.3.3. 'Unaltered' limestone

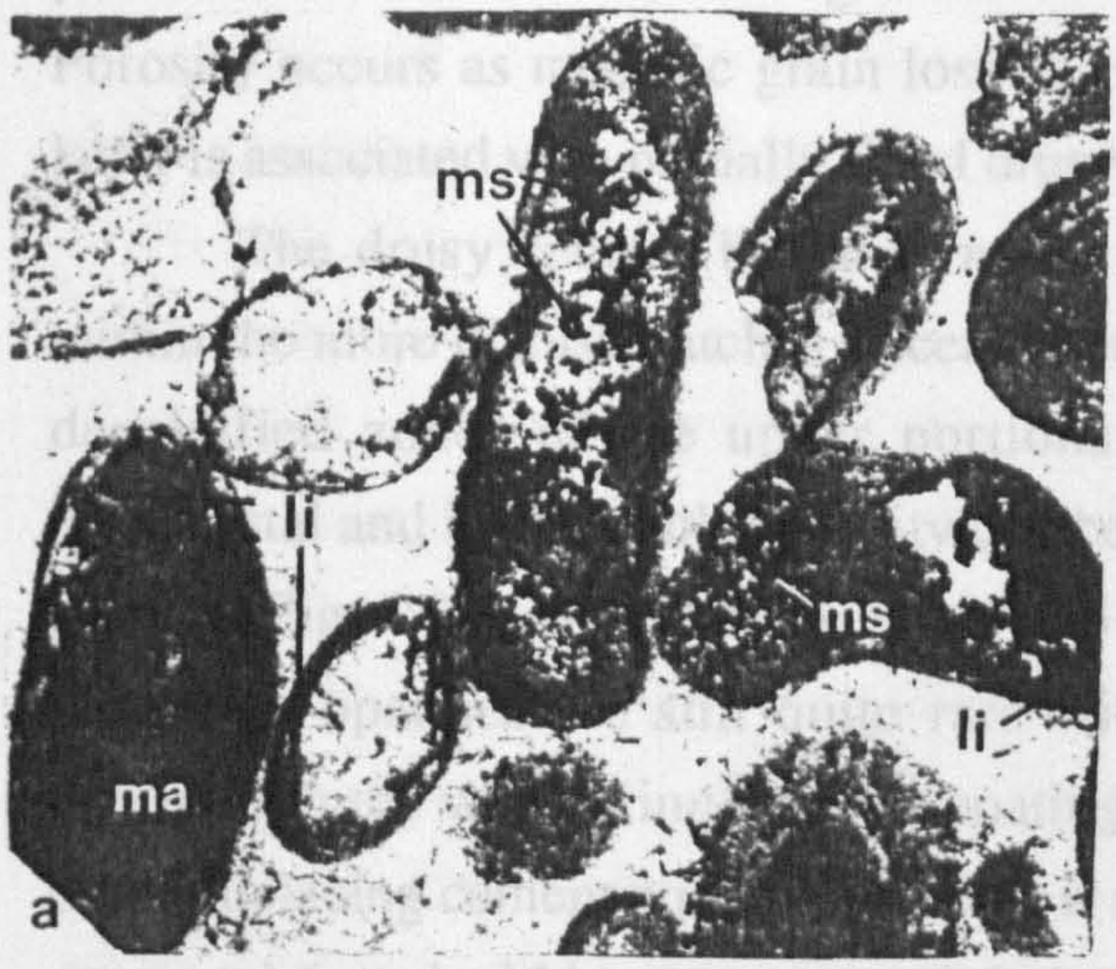
All the specimens taken from the karstic surface possess small inclusions of the original well-cemented limestone, which appear clear or pale yellow in natural daylight. Compositionally, the limestone is made up of sub-equal proportions of bioclasts and ooids, and is a rounded biooosparite (Dunham, 1962). Other allochems make up about 20% of the rock, which is typically a grainstone, although in places the shell fragments are large enough to be classified as a rudstone (for example HQCN3-1, Fig. 4.8b,c) (Folk, 1959, 1962). Micritic cements can form in the marine and freshwater vadose environment as a replacive cement and is common in some palaeokarstic sequences (cf. Davies, 1991). However, micrite is rare (<5% of rock) in the Hornsleasow calcarenites, which are usually cemented by sparite (Fig. 4.8a-e), although, sample HQCN4, contained up to 10% grain-coating micritic matrix in places (Fig. 4.8f). The micrite in this section contains large amounts of ferrihydrite inclusions, and in places forms meniscus contacts between grains (not seen in the illustrated example: Fig. 4.8f), suggesting lithification in the vadose zone.

Quartz is fairly rare and makes up less than 5% of all ooid interiors in the rock, this is in direct contrast to the other samples collected below the hollow (e.g. HQCN1'), which may contain up to 15-20% quartz sand (section 3.5.1). Small inclusions of opaques occur within the sections and are concentrated especially within the weathering rinds of some grains (Fig. 4.8a,c,d,f). These make up around 5-7% of the rock, although this proportion falls to as low as 1-2% in the less altered HQCN4. Relict bedding does not appear in the karstic limestone sections, but they were probably cross-laminated, as this can be observed in specimens of the same bed at lower levels in the section (HQCN1', section 3.5.1).

In terms of cementation history, much of the earlier history of the limestone has been obliterated by karstic and pedogenic overprinting alteration, and the final pervasive ferroan calcite drusy spar, which makes up more than 40% of the rock by modal composition. A few of the specimens do contain relict radial-fibrous and syntaxial overgrowths upon well-preserved echinoid clasts which may have been precipitated in the marine phreatic or vadose zones (section 3.6.1). This early diagenetic cement is typically preserved in areas of the rock which have small interstitial pore-spaces between grains.

The ferroan calcite drusy spar is similar to that seen in other sections from the Chipping Norton limestone. It is typically a pore-filling anhedral drusy cement (Fig. 4.8a-f) and also occurs as a grain replacing cement (Fig. 4.8a,c). Within the well-defined 'peds' of the less-altered mottled or brecciated regions of the karstic limestones (such as HQCN3-1, Fig. 4.8b-d), the cement is largely unaltered and interparticle porosity is very low (1-2%). The cement contains very rare tiny patches and microspherules of opaques which may be weathering products or windborne detrital

Figure 4.8. Photomicrographs of petrographic thin sections of the Chipping Norton Formation unaltered limestone regions underlying the karst surface: (a) section HQCN2-1, plane polarised light; (b) HQC3-1, plane polarised light; (c) HQCN3-1, plane polarised light; (d) HQCN3-1, cross polarised light; (e) HQCN3-1, cross polarised light; (f) HQCN4, cross-polarised light. Field of view 2.3mm. All sections were photographed under blue-filtered light. Abbreviations as in Fig. 3.5, except 'ht' hematite, 'li' limonite, 'ms' microspherules of iron oxides, 'wr' weathering rinds.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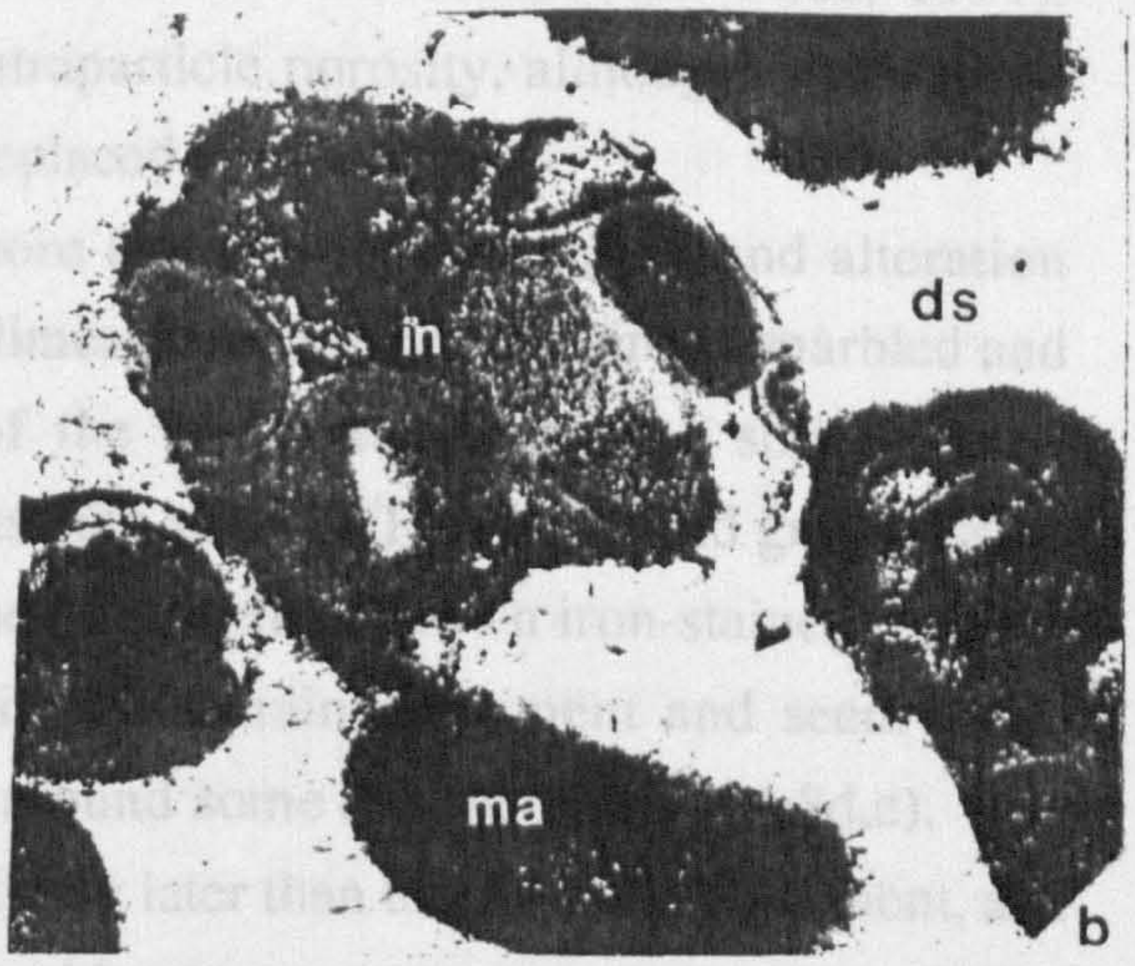
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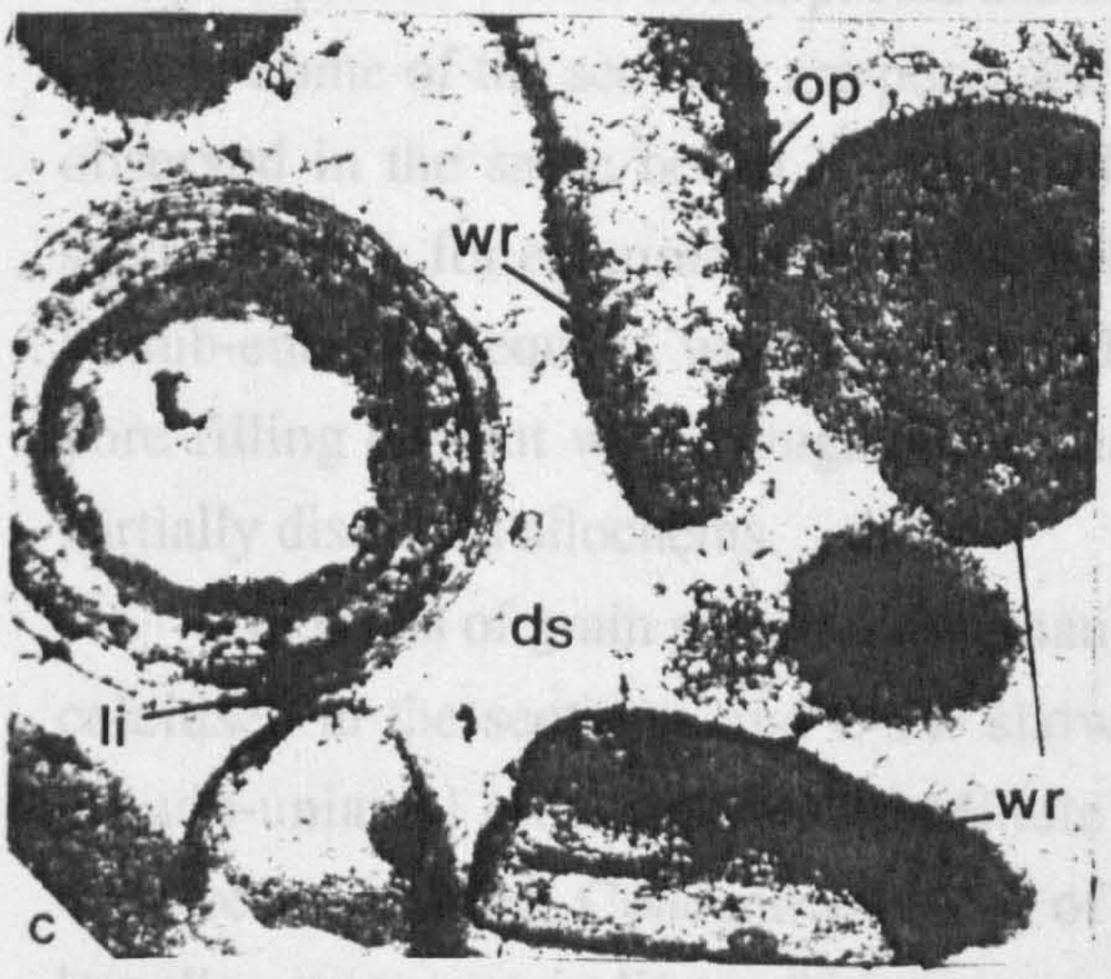
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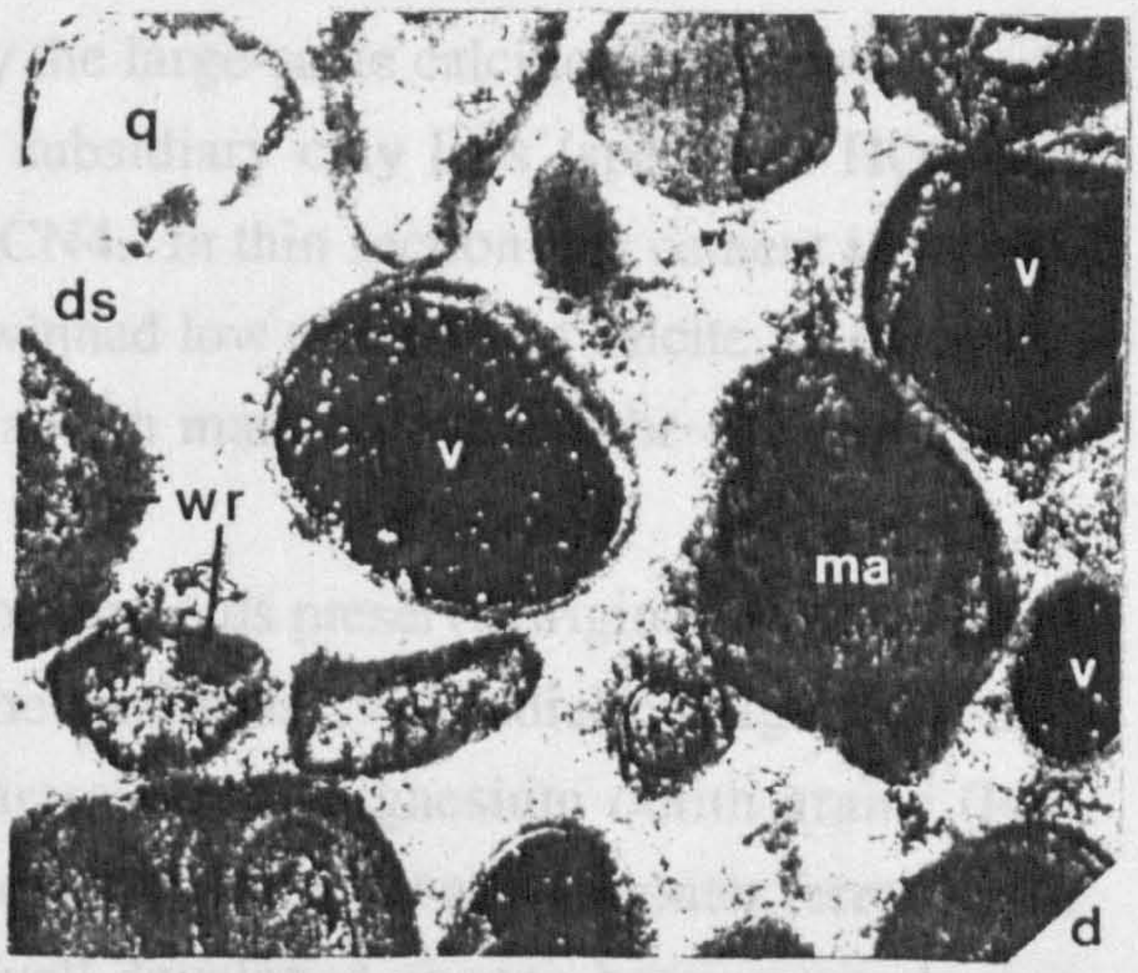
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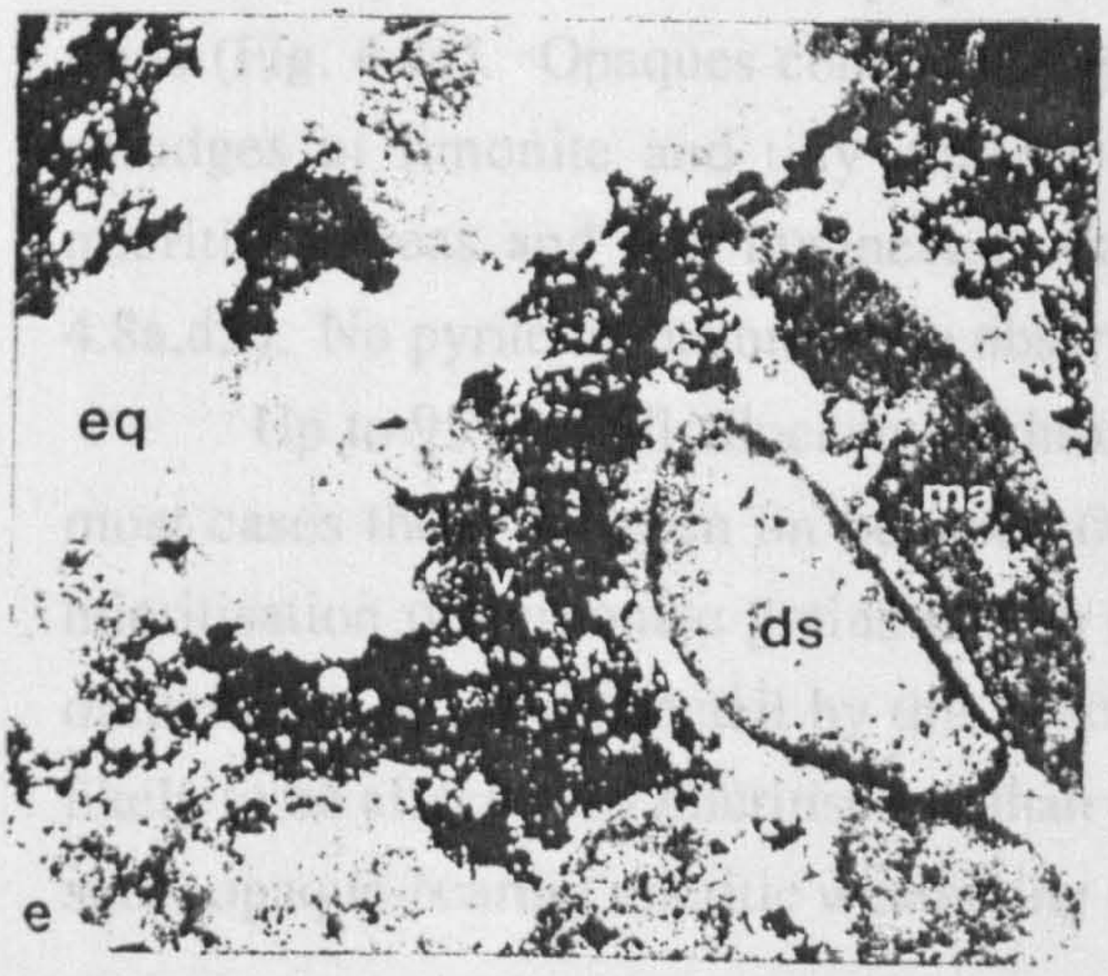
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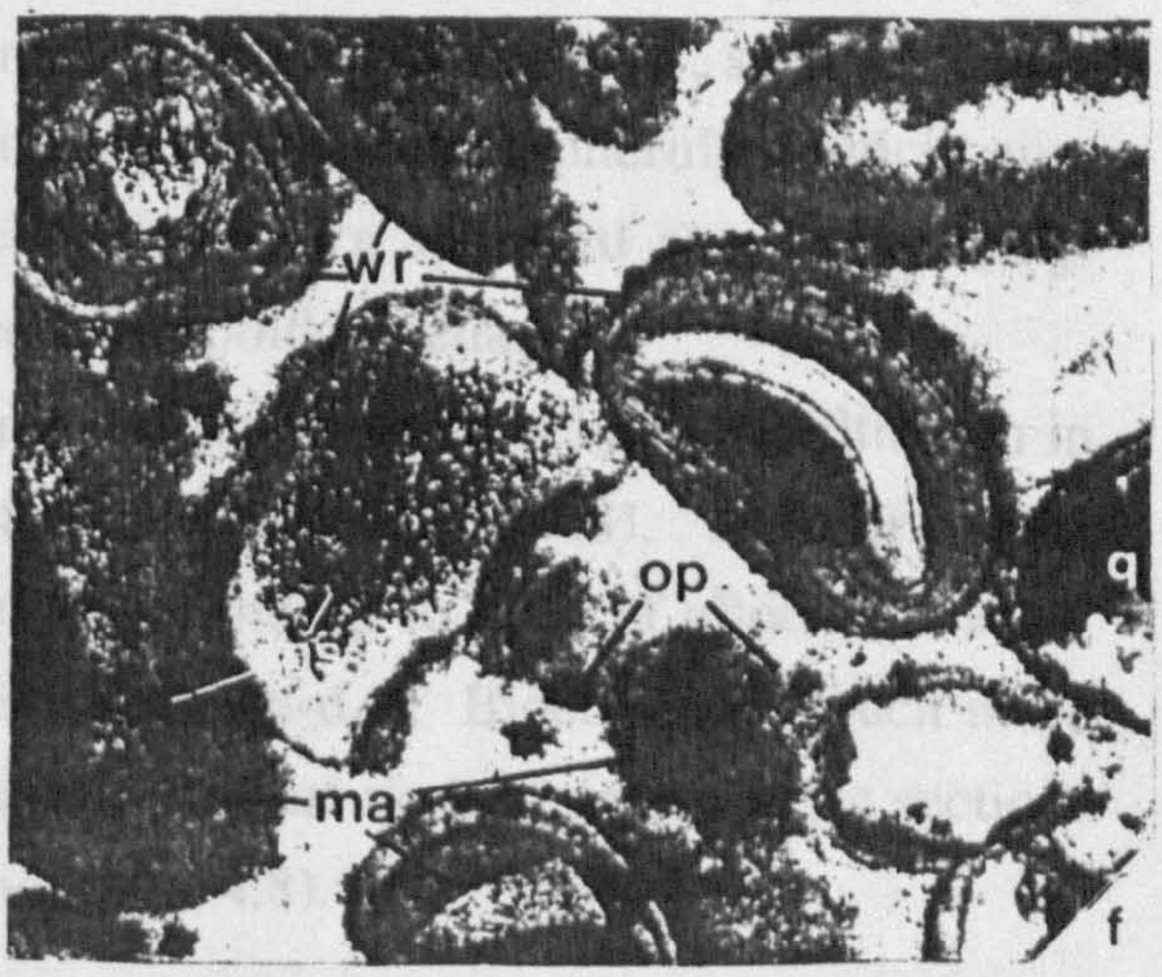
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particles introduced into the grainstone before final cementation (cf. Foos, 1991). Porosity occurs as mouldic grain loss and intraparticle porosity, although much of the latter is associated with partially filled drusy-replaced allochems.

The drusy cement becomes much more affected by dissolution and alteration within the more diffuse patches of cemented limestone occurring within the marbled and decalcified zones of the upper portions of the karst underlying the soils. Here, intercrystal and interparticle porosity can be as high as 15% (Fig. 4.8e), and grain loss up to 20% (Fig. 4.8d). These areas grade into the more yellow-brown iron-stained regions. However, opaques are still quite rare within the remaining cement and seem to be confined to the weathering rinds or coatings around some allochems (Fig. 4.8d,e). The drusy-replacing cement within allochems is slightly later than the interparticle cement, and was precipitated within uncompacted skeletal voids.

Some of the sections are pervaded by the large-scale calcitic veining which was observed in the same bed 0.3m below the subsidiary clay lens (specimen HQCN1': section 3.5.2), for example HQCN2 and HQCN4. In thin section this cement is seen to be sub-euhedral, equant, well-cleaved, and twinned low magnesium calcite. It occurs as pore-filling cement within vugs and veins, and in many places can be seen to dissect partially dissolved allochems.

In terms of grain preservation, many of the ooids preserve original cements or are calcitised in the sections, and 2-5% show the well-preserved radial configuration and pseudo-uniaxial cross, typical of Middle Jurassic low magnesium oolith grains (Fig. 4.8d; section 3.5.3). Concentric zoning of the ooids is present and the outer ferroan-rich lamellae may contain lines of opaques and well-developed opaque-bearing weathering rinds (Fig. 4.8c). Opaques consist of organics and ferric oxyhydrites, such as brown smudges of limonite and tiny reddish-brown goethite microspherules which give micritised areas and low magnesium calcite bioclasts a speckled appearance (Fig. 4.8a,d,f). No pyrite inclusions were observed in the ooid lamellae.

Up to 95% of all allochems exhibit micritic envelope development, although in most cases these are seen on between 60-80% of grains (Fig. 4.8). Total or partial micritisation of carbonate grains is also common, and up to 70% of allochems show obliteration of interior detail by this process (Fig. 4.8a-d,f). Bioclasts are much more likely to be affected by micritisation than ooids, and most skeletal grains in the sections show opaque-bearing micritic weathering rinds (Fig. 4.8).

4.3.4. Yellow-brown friable limonitic limestone

The yellow-brown areas can be described as altered ferruginous oobiosparites, rather than as ironstones. They are compositionally similar to the unaltered regions described above, but have undergone much more alteration and dissolution than the well-cemented regions. In the first instance, the rock is much more friable and the intercrystal and

interparticle porosity has increased to around 15-20% in these areas (Fig. 4.8e). This is associated with much grain loss during preparation of the sections (Fig. 4.8d). No early cements were observed in the altered zones. The sparite cement makes up less than 30% of the rock (Fig. 4.9a,b). Even where the cement remains intact it contains many opaque inclusions, which impart an overall yellow appearance to the sparite in hand specimen (Fig. 4.6a). The dissolution of the case-hardening cement is complemented by an increase in interstitial ?replacive dirty-brown micrite, which forms up to 15% of the matrix and contains large diffuse opaque smudges and granules (Fig. 4.9b).

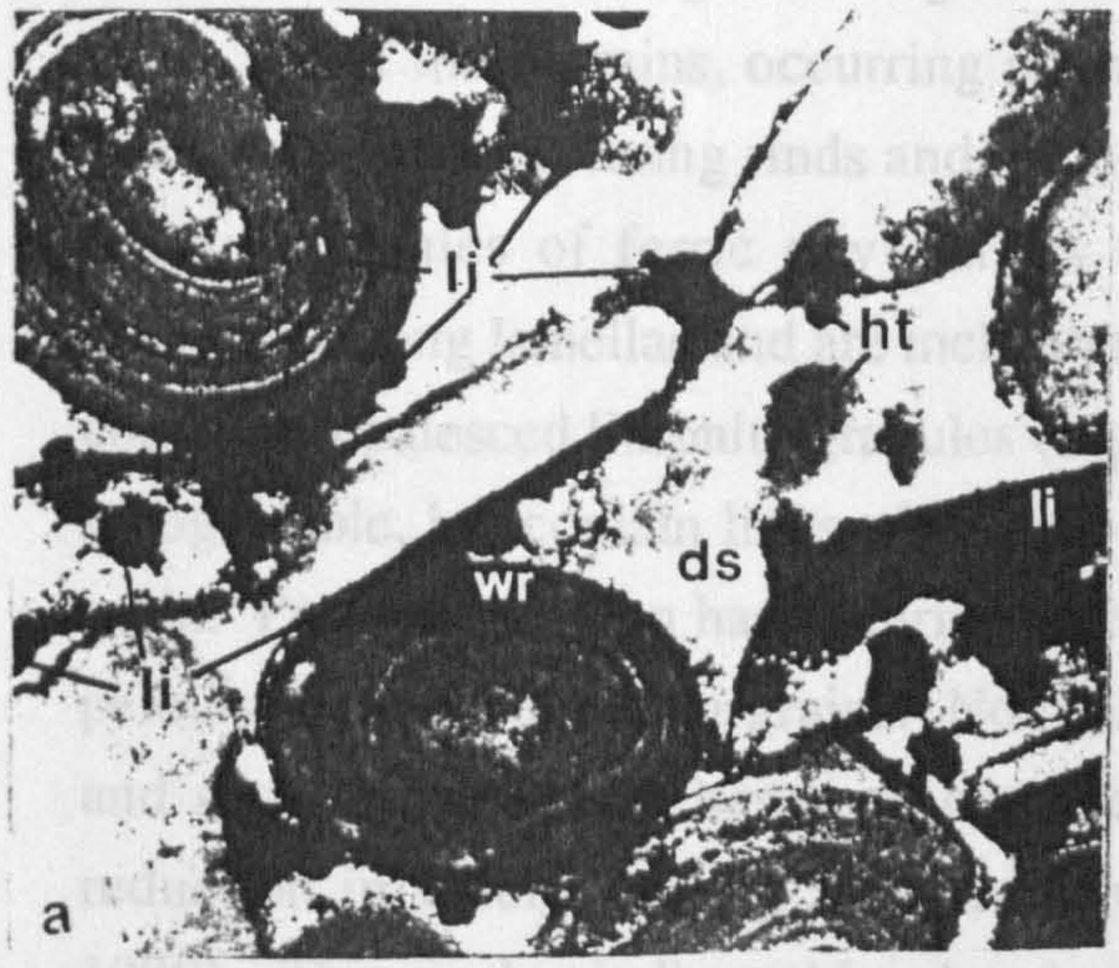
The preservation of the grains is similar to that in the unaltered areas. Heavily eroded brown weathering rinds occur on the majority of allochems including ooids and these are inclusion-rich (Fig. 4.9a,b). The weathering rinds are particularly visible upon bioclast grains such as echinoids and oysters, and are composed of a dirty brown micrite which contains many large granules of limonite (Fig. 4.9b). In addition, the higher proportions of non-ooid quartz in these areas (up to 15% in HQCN2-1) suggests total dissolution and consolidation of superficial oolith grains. The interior of unmicritised low magnesium calcite grains such as echinoids and oysters show lines of red and reddish brown ferric oxyhydrite and iron oxide microspherules (Fig. 4.9a,b).

4.3.5. Dark reddish-brown decalcified surface ironstone

The pervasive decalcified region underlying the soil units within the hollow is best seen in sample HQCN3-2. In thin section the rock is compositionally a calcareous haematitic ironstone (Fig. 4.9c,d), and bears only a superficial resemblance to the underlying karstic oolites. The rock is composed of 30-40% recognisable allochem grains and 5-10% quartz grains set within a matrix composed of deep red aggregated haematite and black-brown smudges of limonite, possible coalified organics and other oxides (?titanium and manganese oxides) (Fig. 4.9c,d). Haematite is strongly coloured in plane polarised light, but is not entirely opaque (Fig. 4.9c), and in cross polars it exhibits low order birefringence colours (Fig. 4.9d), and is therefore easily distinguished from opaque limonite and goethite. The haematite occurs as a pore-lining 'cement' and is composed of aggregated grain-coating spherules (Fig. 4.9c,d). Sparite cement makes up less than 5% of the pore space (Fig. 4.9d) and micrite about the same proportion. The decalcified region is particularly friable, with high interparticle, intraparticle and intercrystal porosity (Fig. 4.9c,d). During preparation the loosely cemented ironstone lost up to 70% of grains. All grains exhibit haematite or opaque coatings, and the decalcified area can be considered to have a 'coat and bridge' structure (FitzPatrick, 1993), similar to the ferruginous regolith typical of heavily weathered lateritic *'terra-rossa'* soils (section 4.3.8).

The allochems are similar in composition to those within the yellow-brown ferruginous limestones. Most recognisable allochems show well-developed micritic

Figure 4.9. Photomicrographs of petrographic thin sections of the Chipping Norton Formation altered limestone regions underlying the karst surface: (a) section HQCN3-2, cross polarised light; (b) HQCN2-1, plane polarised light; (c) HQCN3-2, plane polarised light; (d) HQCN3-2, cross polarised light; (e) HQCN3-1, cross polarised light; (f) HQCN3-1, plane polarised light. Field of view 2.3mm. All sections were photographed under blue-filtered light. Abbreviations as in Figs. 3.5 and 4.8.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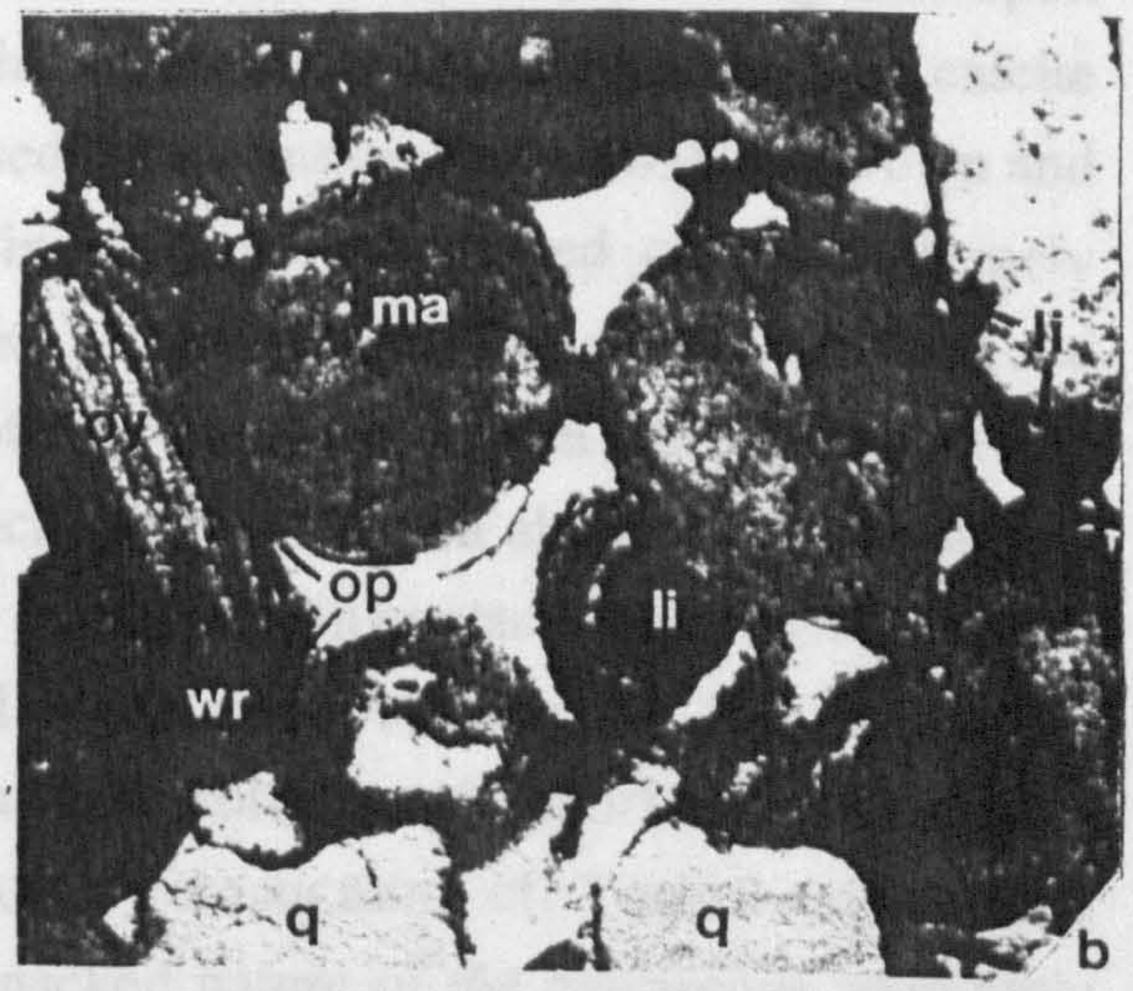
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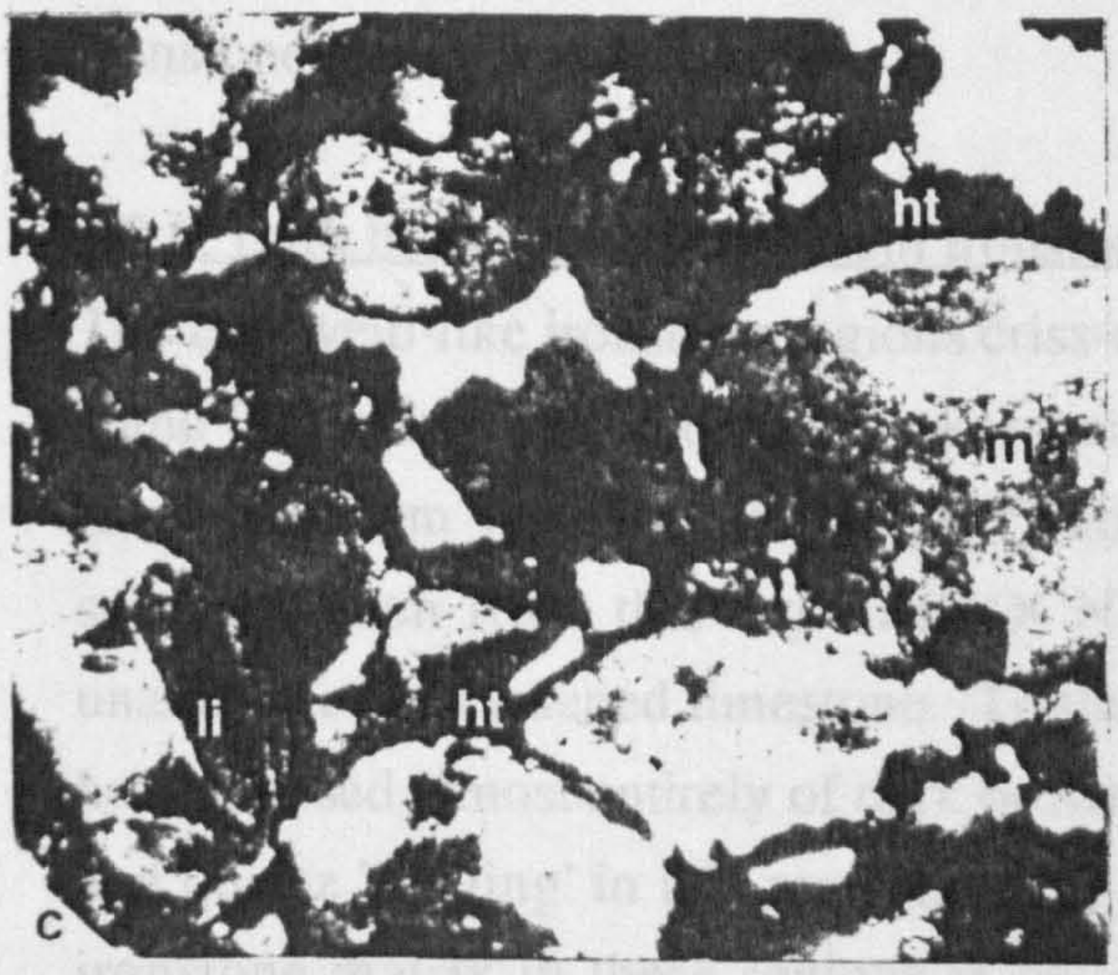
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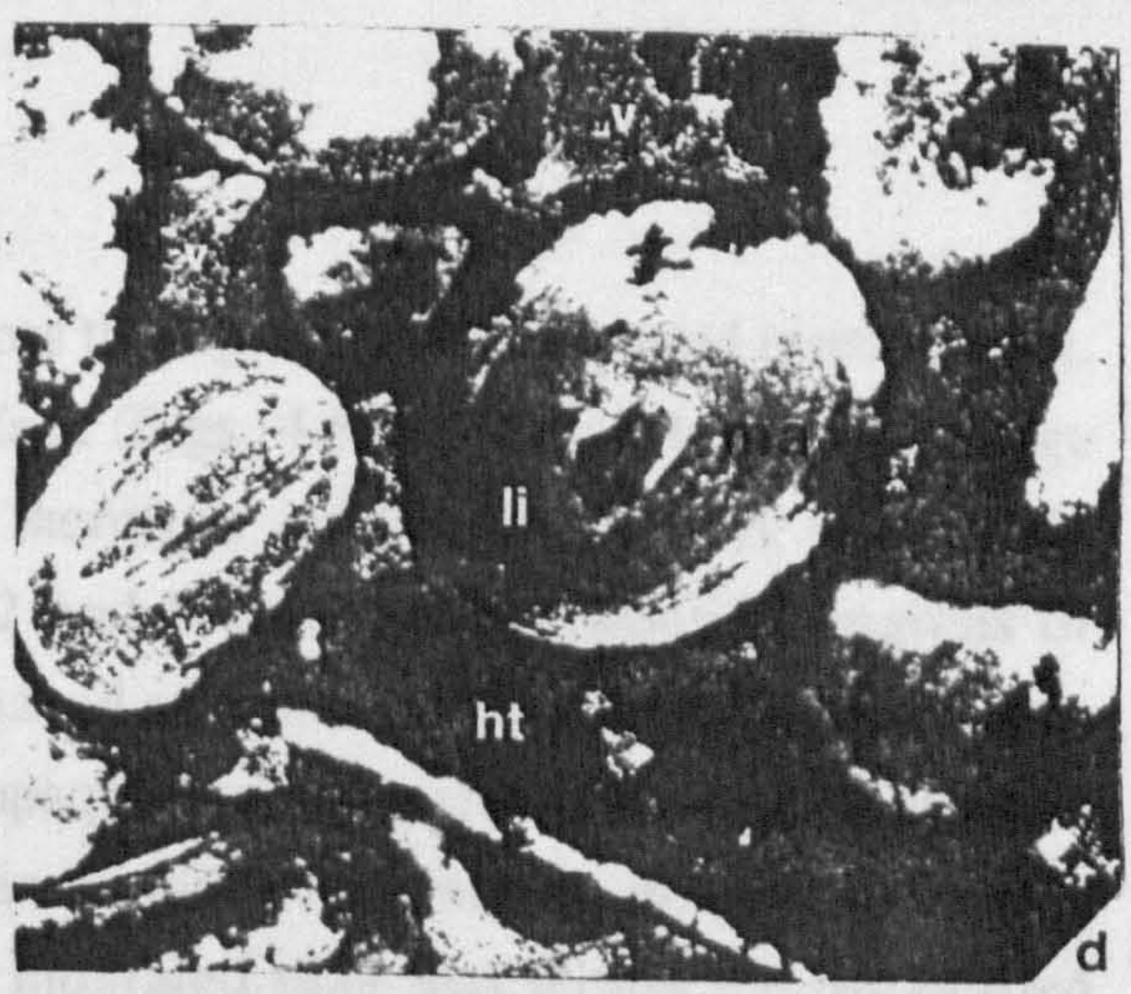
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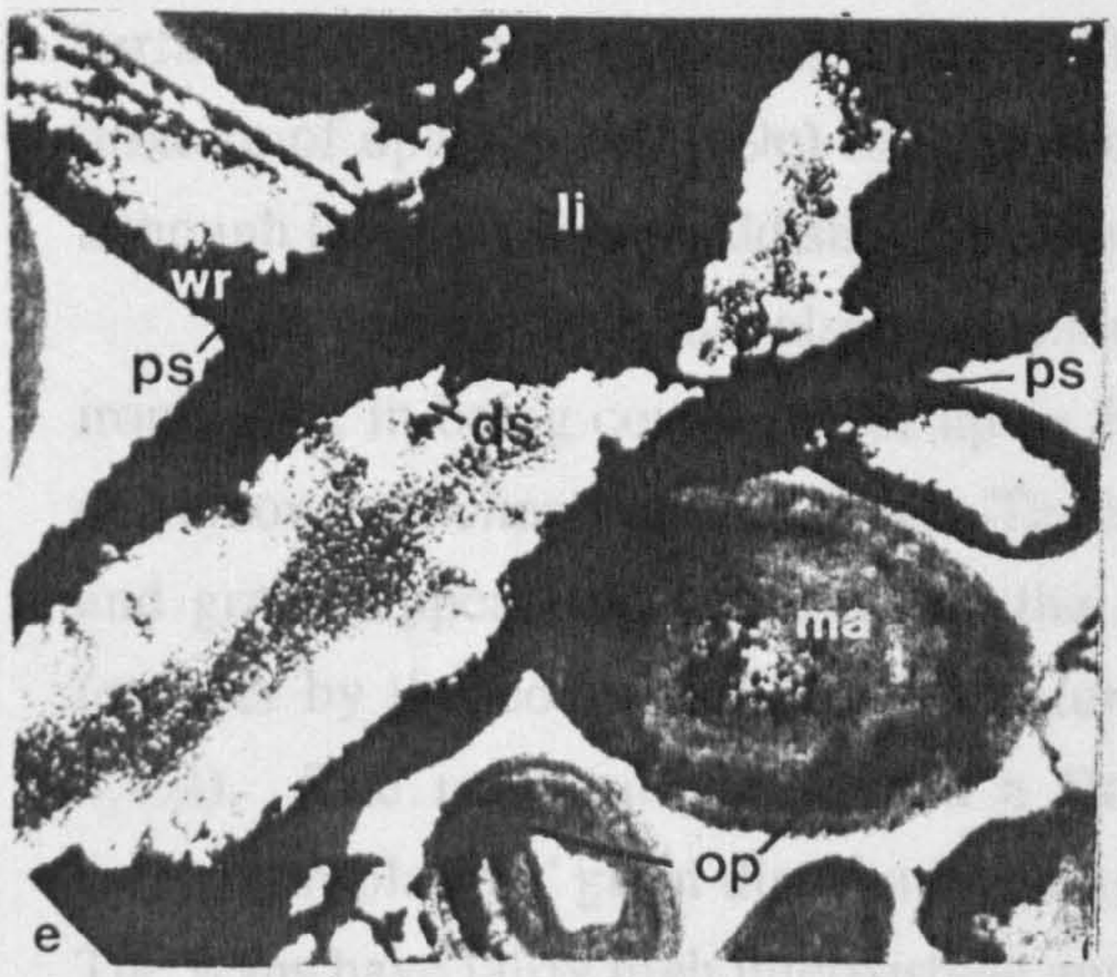
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weathering rinds, although the original micritic envelopes can only be recognised upon about 5% of these grains, occurring upon the better preserved low magnesium calcite bioclasts. The weathering rinds and micritised allochems are rich in reddish-brown and opaque granules of ferric oxyhydrites (Fig. 4.9c,d). Calcitised ooids are largely micritised along lamellae and are inclusion rich. The outer lamellae are impregnated and coated by coalesced limonite granules (Fig. 4.9d). Low magnesium calcite bioclasts are recognisable, but contain lines of opaque inclusions developed along cracks within the grain. Pressure solution has occurred along grain-to-grain contacts associated with soft peloids pushed into harder grains. Non-ooid quartz makes up around 5-10% of the rock and suggests that there has been considerable dissolution of superficial ooids and a reduction in volume during alteration ('vadose compaction': cf. Tucker and Wright, 1990). This is also indicated by the close-packed nature of the calcareous haematitic ironstone areas (Fig. 4.9c,d).

4.3.6. Dark brown decalcified vein ironstone

The thin, vein-like ironstone regions criss-cross through the well cemented mottled zone, some 0.1 metres below the palaeokarst surface (Figs. 4.6a & 4.7). These veins range from 0.5-2mm in width (Fig. 4.9e,f) and therefore are clearly defined, in the thin sections taken from this region (HQCN2.2 and HQCN3.1), surrounding regions of unaltered case-hardened limestone. To the naked eye, these veins or stylolites appear to be composed almost entirely of dark brown opaques, with occasional calcareous material and quartz 'floating' in this matrix. They can be termed calcareous ironstones, but the ironstone matrix in these regions is better indurated than that within the decalcified surface ironstone and forms a cement. The ironstone cement is largely composed of a mixture of opaque (Fig. 4.9e) or deep brown massive limonite and micrite (Fig. 4.9f), although there are some reddish-brown iron oxide or goethite inclusions (Fig. 4.9e).

These ironstones are also dissimilar in composition to the upper friable haematitic ironstones, in being composed of up to 50% limonite 'cement' and as little as 10-15% calcareous allochems (Fig. 4.9e,f). Texturally, the ironstone matrix is fairly massive, and grains appear to be floating in the matrix (Fig. 4.9f), rather than merely stuck together by the 'coat and bridge' structure exhibited by the upper ironstones (section 4.4.4). The rock can be termed a floatstone (Dunham, 1962) or as having an 'agglomeroplasmic' grain distribution in terms of soils classification (Retallack, 1990). The veins have fairly high interparticle and intercrystal porosity caused by dissolution of the carbonate cements (Fig. 4.9e), although the grain loss during slide preparation is much lower (at around 10-15%) than in the poorly cemented coated-grain surface region.

Most of the allochems are peloidal or remnant partially micritised bioclasts, and ooids make up only about 5% of the total grain content. Non-ooid quartz is quite common in these decalcified areas and specimen HQCN2.2 contains around 10-15%,

compared to less than 5% in the surrounding limestone occurring as oolith cores. Original and replacive cement and matrix is extremely rare within these vein regions, with less than 5% micrite and sparite in both sections (Fig. 4.9e). Where rare drusy sparite occurs the cement is full of dark brown ferric oxyhydrite opaque inclusions.

Clasts within the solutional boundaries are quite poorly preserved in comparison with those in the surrounding limestone (Fig. 4.9f). Of the bioclasts only 1-2% are preserved well enough for clear identification and these tend to have been originally composed of low magnesium calcite (e.g. oyster fragments) but show dissolution along cracks and are surrounded by well-developed opaque-rich micritic weathering rinds. The rest of the bioclasts are almost totally micritised and exhibit ghost textures. Drusy-replaced bioclasts are not distinguishable from the regions of relict sparite cement. Pelletal and micritised peloids contain abundant inclusions of reddish-brown goethite, and all allochems display inclusion filled dirty-brown micritic weathering rinds. Ooids are calcitised, with abundant opaque inclusions within the outer lamellae. Even the quartz grains exhibit the first stages of weathering in displaying grain-cloudiness and dissolutional along cracks and around the edges (FitzPatrick, 1993).

4.3.7. The pebble-like clasts associated with the karst

The two pebble-like clasts sectioned and examined are compositionally different limestones from the bed which has been weathered by karst processes. However, both possess similar weathering features typical of the karstic limestones and are considered to have undergone erosion within the doline prior to soil deposition. The clasts exhibit a heavily iron-stained outer limonitic crust (Fig. 4.6b,c), which is fairly well indurated. This crust was examined in the section taken from specimen HQP1 (Fig. 4.10b). The interior of the clasts also appear to be non-corroded and well cemented.

Compositionally, HQP1 is similar to beds occurring elsewhere in the Chipping Norton Formation and was probably derived from a bed above or around the hollow during karstic planation of the landscape. It is a well cemented equigranular calcarenite and the interior of the clast appears dark browny-grey in thin section. The clast is texturally a wackestone (Dunham, 1962), containing sub-equal proportions of micrite, which appears to be primary carbonate mud (which imparts the colour in thin section) and sparite cement (Fig. 4.10a), and is classified as a poorly washed well sorted oopelsparite (Folk, 1959, 1962). The rock also contains up to 5% quartz occurring as ooid cores and 5-10% opaque inclusions disseminated throughout the micritic regions (Fig. 4.10a).

HQP1 is cemented by a pore-filling drusy sparite cement which may be an early marine, beachrock or case-hardening cement. Porosity is quite low in this section, and only 5-10% of clasts were lost during preparation. The preservation of the allochems in the specimen is similar to those preserved in the unaltered parts of the karstic limestones at Hornsleasow. Allochems often show micritic envelopes and up to 50% of all bioclasts

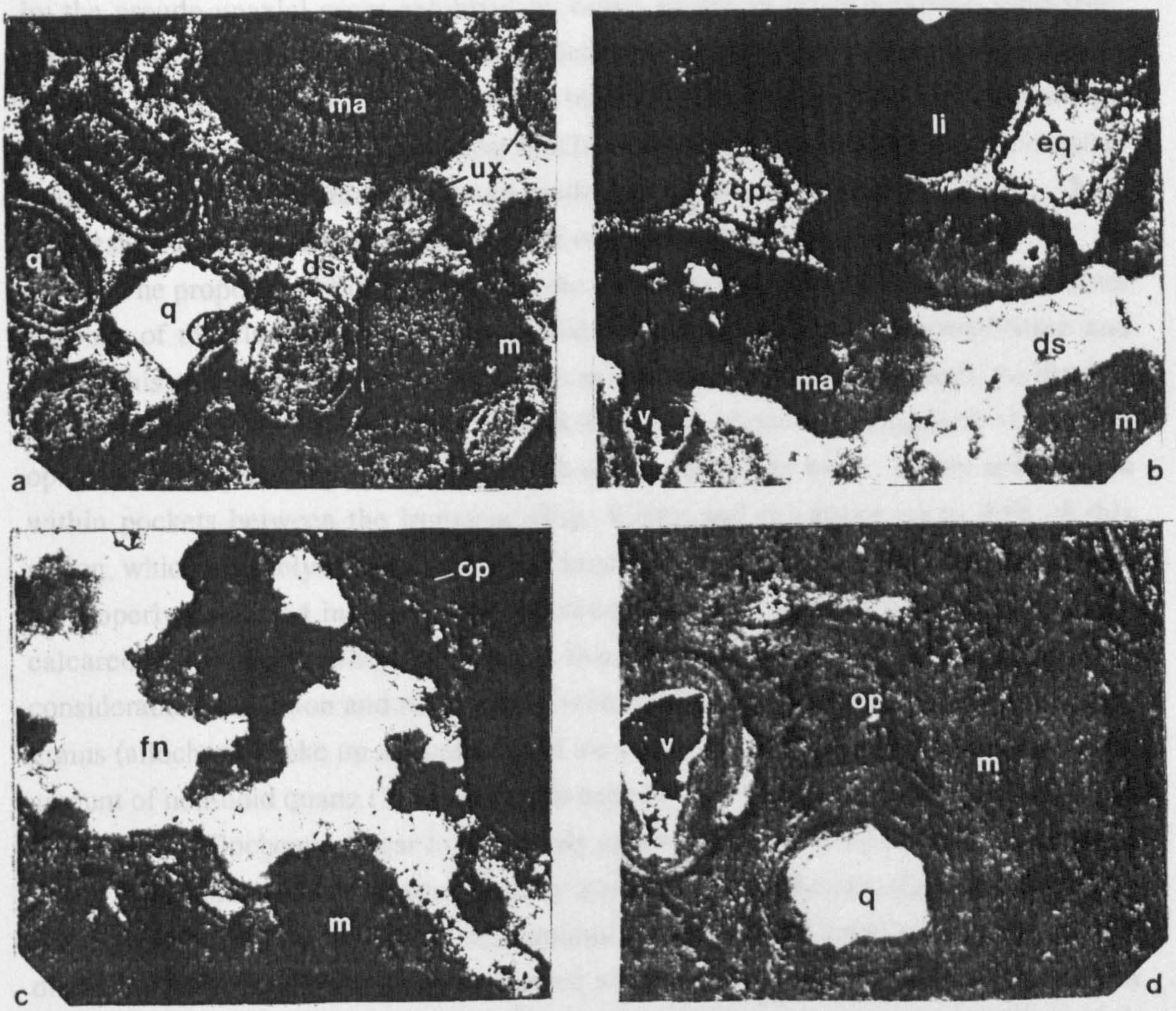


Figure 4.10. Photomicrographs of petrographic thin sections of the altered limestone pebble-like clasts sampled from above the karst surface: (a) section HQP1, cross polarised light; (b) HQP1, cross polarised light; (c) HQP2, plane polarised light; (d) HQP2, cross polarised light Field of view 2.3mm. All sections were photographed under blue-filtered light. Abbreviations as in Figs. 3.5 and 4.8.

have been partially micritised in HQP1 (Fig. 4.10a,b). The ooids in the section are calcitised and exhibit well-preserved compositional zoning and radial structure illustrated by the pseudo-unaxial cross exhibited by oolith grains in cross-polarised light (Fig. 4.10a). The outer ferroan rich lamellae preferentially display oxidation products such as limonite and goethite (Fig. 4.10a,d). A few bioclasts have been completely replaced by a drusy spar in section HQP1, and most bioclasts show total micritisation or incomplete micritic weathering rinds associated with opaque coatings (Fig. 4.10a,d). Low magnesium calcite bioclasts contain lines of opaque inclusions (Fig. 4.10a).

The proportion of opaques within the section strongly increases toward the outer portions of the clast where a thin limonitic 'skin' is developed. In composition and texture this crust is similar to the vein-like areas of dissolution which occur in the mottled regions of the palaeokarst (Figs. 4.6a & 4.9e,f). The matrix is composed of massive opaque limonite (Fig. 4.10b), with less than 5% carbonate mud. Drusy spar occurs within pockets between the ironstone (Fig. 4.10b) and can make up to 20% of this region, which probably accounts for the induration of the crust. The region is too thin to be properly classified in terms of composition or texture, but is best described as a calcareous ironstone, where grains are floating within a limonite matrix. Again considerable dissolution and reduction in volume is indicated by the substantial loss of grains (allochems make up around 10% of the crust) and complimentary increase in the amount of non-ooid quartz (10-15%) in this region.

Most allochems appear to be heavily sparmicritised toward the outer crust (Fig. 4.10b). The sparite cement appears heavily iron stained and contains abundant inclusions (Fig. 4.10b). The structure of the limonitic matrix goes from a coat and bridge distribution in the interior to a much more massive and pervasive distribution over a distance of less than one millimetre (Fig. 4.10b). Associated with this is a fairly high degree of preparatory grain loss in the outer-most iron-rich crust.

The second clast, HQP2, is quite unlike any of the limestone beds seen within the Chipping Norton Formation at Hornsleasow Quarry. Overlying contamination of the clay lens cannot be ruled out entirely, as cambering fissures may introduce exotic material, but contamination during excavation is extremely unlikely as this area was very carefully sampled by M.J. Simms (pers. comm., 1992). However, the similar weathering features and limonitic crust do provide reasonable proof that this is not the case. HQP2 is a calcilutite, and in both hand specimen and thin section large irregular pore spaces, called 'fenestrae' or 'birdseyes', have been observed within the dark green carbonate mud matrix (Fig. 4.10c). There are sparse allochems (between 5-10%) composed of essentially micritised ooids, bioclasts and peloids (Fig. 4.10d). Quartz grains are extremely rare and occur only within oolith cores (Fig. 4.10d) and 'strings' of opaques are disseminated throughout the matrix (Fig. 4.10d) and in some places appear to define a rough lamination (Fig. 4.10c). Some of the fenestrae part along the laminae

and are probably 'laminoid fenestrae'. Section HQP2 is extremely friable, the initial fenestral porosity is largely infilled by a drusy slightly ferroan calcite cement (Fig. 4.10c), but the loose cementation of the carbonate muddy matrix means that there is much secondary loss of grains during preparation (Fig. 4.10d).

4.3.8. Interpretation of petrographic features of the palaeokarst

The calcarenites of the Chipping Norton Formation below the karst show varying degrees of decalcification and are in varying states of cementation. The karst specimens were taken from the top 0.3m of the bed which has undergone karstification, and exhibit segregated zones of chemical alteration from the surface downwards. The descriptions of features seen in the petrographic study of the limestones are now considered together as part of a karstic weathering profile below the paleosol at Hornsleasow.

The weathering profile begins about 1-1.5m below the surface of the palaeokarst were thin dissolution veins of limonite-rich ironstone are seen transecting the well-cemented sparry calcarenite. This area is known as the mottled area of dissolution (Fig. 4.7) and is analogous to the platy or nodular weathering zone of modern karstified limestones described by Hine *et al.* (1988) underlying coastal marshes in northwest Florida and the mottled 'Ck' horizon beneath lateritic soils in the Bahamas (Foos, 1991).

The pore-lining cement was removed from the Chipping Norton Formation grainstones in the freshwater phreatic environment (section 3.6.1) prior to actual karstification. The later drusy cement is not affected by dissolution, but becomes so within the more diffuse patches of cemented limestone occurring within the marbled and decalcified zones of the upper portions of the karst (Figs. 4.6a & 4.7), underlying the soils. The drusy cement was probably precipitated as low magnesium calcite in the meteoric environment (section 3.6.1) and could be associated with the case-hardening cement, although precipitation as an early marine or late burial cement cannot be ruled out (section 3.6.1). If the latter case was true, then the opaque inclusions were probably picked up by the cement during precipitation.

However, sparry calcite cement is laid down at present in the freshwater vadose environment of humid weathering karstic regions (cf. Foos, 1991) and there is evidence that the limestones were at least lithified and cemented at the surface prior to sub-aerial karstification. This is the case-hardening cementation of Trudgill (1985). Brecciation of the limestone has occurred below the soils, leaving a network of dissolutional 'veins' of limonitic ironstone which pervade the mottled zone of the profile and seem to follow the regular jointing pattern, as well as occurring along bedding planes. Also, the large-scale karstic features on the surface of the limestone, could only have been formed upon an indurated profile. This suggests that the limestone was at least partially indurated prior to dissolution during karstification and pedogenesis.

Preservation of the allochems within the well-cemented limestone areas gives a much closer insight into the alteration of the limestone beneath the palaeokarstic surface. For it is these grains which appear to have undergone a protracted period of weathering and alteration. Opaque weathering products (iron oxyhydrites) are seen as inclusions or coatings upon many of the grains. Also most allochems exhibit strong micritisation. Trudgill (1985) has stated that a high degree of micritisation is associated with biochemical erosion upon present-day coastal karstic areas, and other authors have since corroborated this veiw (Hine *et al.*, 1988; Foos, 1991). Even well-cemented coastal beachrock can be heavily eroded by encrusting, grazing and boring organisms which inhabit these areas (Tucker and Wright, 1990). The micritisation and erosion of allochems could have taken place in the littoral zone prior to inland karstic weathering, although cementation would have limited this effect.

Another likely origin of the micritisation of both grains and sparry cement has been outlined by Davies (1991) in his description of micritised grains occurring within a palaeokarstic sequence of Dinantian limestones upon Anglesey. He observed micritisation of grains up to two metres below the palaeokarstic surface, and noted an increase in degree and amount of grains micritised toward the top mottled parts of the profile. He suggests that the micritisation takes place by degrading recrystallisation beneath the case-hardened palaeokarst surface (Davies, 1991). A replacive origin would seem to be supported within the Hornsleasow limestones, as the weathering rinds show fairly sharp outer edges (Figs. 4.8a,f and 4.9a,b) and retain their original shapes. The process of transformation is not well established, but seems to be caused by a combination of dissolution of the original carbonate and concomitant re-precipitation of micrite by percolating porewaters enriched in organic chemicals released from the bacterial decay of organic matter (Kahle, 1977). The organic-rich porewaters are also thought to inhibit growth of calcite crystals beyond the microspar stage. There is abundant plant matter within the overlying paleosols and upon the karst surface at Hornsleasow which could provide potential porewater fluids. The process is known as 'sparmicritisation' (Kahle, 1977).

Davies (1991) also describes associated micritic features called 'calcrete ooids' and 'calcrete peloids' in the Dinantian karst profiles and these grains also occur in Quaternary calcrete profiles (Read, 1974, Harrison, 1977). The 'calcrete ooids' are usually composed of brown micritic material surrounding a nucleus. In simple terms these and the calcrete peloids are replaced allochems, but other processes can result in their formation within karstified limestones and the soils which overly them. For instance, the peloids may be faecal pellets of soil organisms (Wright, 1986, 1987) or tiny carbonate nodules (section 5.9.3), and the ooids may form from growth of a micritic cement around the nucleus, within the soil matrix (carbonate concretions, section 5.9.3) or can occur as displacive features within the limestone itself. These occasionally form

surface crusts, which resemble ooid or pelloidal grainstones (Davies, 1991). Those which are precipitated in the soils as nodules may also occur in cracks and fissures within the host limestone, although they may enclose grains of soil constituents, such as clay or terrigenous quartz (Davies, 1991). Although they originate through pedogenic processes, these grains are virtually indistinguishable from secondarily micritised normal marine allochems and therefore their identification within the Hornsleasow profile is almost impossible.

The limonite ironstone veins which cut across the limestone in the mottled and marbled regions are interpreted as discrete solutional pathways. The reduced proportions of carbonate grains and cement in the veins and the more than 100% increase in quartz content, would suggest that there has been total dissolution of grains and a substantial decrease in rock volume along these solution pathways (Fig. 4.9e). Clasts within the solutional boundaries are extremely poorly preserved and suggest that the regions were subjected to prolonged dissolution and erosion. This also is illustrated by the preferential dissolution of dissected grains in these stylolitic regions (Fig. 4.9e,f). Fig. 4.9e shows two such bioclasts which have been dissolved preferentially by a thin limonite-vein which dissects and brings together along a bedding plane, a sparite cemented rudstone in the upper left part of the photomicrograph and grainstone beds in the lower right portion. Other allochems within the unaltered limestones surrounding the vein show preferential dissolution and show an increased content of opaque inclusions and coatings next to the ironstone veins (Fig. 4.9e,f).

The brecciation of the host limestones and formation of these solutional pathways occurred during karst formation and perhaps even underlying the paleosols. Contemporary brecciation of limestones within soil covered karst profiles is caused by alternating periods of wetting and drying, which allow cracks to form by shrinkage and desiccation followed by porewater infiltration and dissolution. These features are common in other palaeokarstic sequences (cf. Davies, 1991) and the mottled region can be considered to be a saprolitic or 'C' horizon of weathered rock underlying the soils. Further dissolution along conduits in the subsurface freshwater environment eventually leads to the development of cave systems (Tucker and Wright, 1990), but this has not happened at Hornsleasow.

The veins are of similar proportions to the clay and iron oxide filled solution voids described by Foos (1991), which occur in the karstified regolith (Ck) horizon underlying modern lateritic mineral soils upon Eleuthera island in the Bahamas. She interprets the voids as being formed or at least utilised by plant roots, before they were closed by an infill of weathered products derived from the overlying soils. The Hornsleasow infill comprises massive limonite, aggregated hematite, micrite and other opaques and suggests derivation from the surface decalcified ironstone region. As a cement it has formed an indurated barrier to further dissolution along these pathways.

Foos (1991) has suggested that such a barrier to dissolution would allow diffusion of the porewaters into the surrounding limestones causing discrete solution and precipitation of carbonates. Micritisation of grains and deposition of clay-sized particles of opaques as pore-lining cements or grain coatings would occur in the 'unaltered' limestone surrounding the veins (Fig. 4.9e).

The discrete dissolution along vein-like conduits of the mottled zone gives way to the marbled area about 0.5-1.5m below the surface (Fig. 4.7), where more pervasive dissolution seems to have taken place. This is made up of areas of decalcified haematitic-ironstone, yellow-brown limonite-stained altered limestone and smaller regions (typically 10mm^2) of creamy-white 'well-cemented' limestone (Figs. 4.6a & 4.7). The boundaries of these areas are quite diffuse, giving the rock a 'marble-like' appearance. The yellow-brown ferruginous oobiosparites are compositionally similar to the unaltered regions described above, but have undergone much more alteration and dissolution. No early marine cements were observed in the altered zones and sparite cement is also rare and inclusion-rich. Interstitial dirty-brown opaque-rich micrite forms much of the cement. It is likely that some of the micrite consists of compacted and partially dissolved peloidal grains, as many of these grains appear squashed and have been forced into harder grains at point-to-point contacts (Fig. 4.9b). The spacing of the grains is much closer than in the original limestone and this probably is a result of post-karst compaction of the poorly cemented grainstone (Fig. 4.9a,b). However, grain-to-grain contact solution and deformation has also been described as a product of undersaturated meteoric vadose water dissolution and is called 'vadose compaction' (Tucker and Wright, 1990). Therefore, much of this grain overpacking probably occurred during karstification. The preservation of the grains in these area is similar to that in the unaltered limestone and again most of the grains have undergone coastal or pedogenic micritisation (Fig. 4.9a,b).

At the surface (up to 50-70mm) the rock has undergone maximum decalcification by freshwater vadose weathering and has left a friable deposit in which remnant grains (highly altered allochems and quartz grains) are loosely cemented by a reddish-brown haematite-rich ironstone cement. Remnant diffuse areas of better cemented, but altered limestone also occur in this region and these may occur as prominent surface projections, or strung out along possible ghost bedding-planes (Figs. 4.6a & 4.7). The haematitic ironstone surface zone is similar to the red *'terra rossa'* clays which have been recognised in Dinantian, Jurassic and Pleistocene paleosols (Riding & Wright, 1981; Wright & Wilson, 1987; Ruhe *et al.*, 1961; Davies, 1991) and in modern lateritic weathering environments (Esteban and Klappa, 1983; Foos, 1991).

These *terra rossa* clays are characteristic of karst terrains, and form from extensive surface limestone dissolution under humid conditions and the accumulation of a thin cover (usually less than 0.2m) of insoluble residues at the surface. In most cases the lateritic soil comprises deeply red-stained aluminosilicate clay, aggregated iron oxides,

manganese oxides and titanium oxides, which in the modern soils, have been derived as detrital particles (e.g. windborne volcanic or desert dust: Foos, 1991). In the case of the Hornsleasow profile, the detrital residue is almost completely haematitic. Mixed in with this detrital material is the unweathered residue derived from the underlying limestones, such as oolite intraclasts, allochems and ooid interiors (quartz). Modern *terra rossa* soils may be rootleted and bioturbated, and are overlain by a thin leaf-litter horizon (Foos, 1991). No bioturbation is indicated in the extremely thin *terra rossa* soil at Hornsleasow, but carbonised plant material has been found at the base of the hollow (section 5.1). Foos (1991) has suggested that *terra rossa* formation occurs in a tropical humid climate subjected to seasonal changes in moisture levels.

The two pebble-like clasts sectioned and examined are compositionally different limestones from the bed which has been weathered by karst processes. However, both possess similar weathering features typical of the karstic limestones and are considered to have undergone erosion within the doline prior to soil deposition. Compositionally, HQP1 is similar to beds occurring elsewhere in the Chipping Norton Formation and was probably derived from a bed above or around the hollow during karstic planation of the landscape. The second clast, HQP2, is quite unlike any of the limestone beds seen within the Chipping Norton Formation at Hornsleasow Quarry. The clast is composed of a laminated lime mudstone. It has very few grains, but contains spar-filled birds-eye fenestrae, which are common desiccation features in modern carbonate mud sediments exposed in the intertidal and supratidal zones (Tucker & Wright, 1990). However, birdseyes can also be found in subtidal hardground sediments (Shinn, 1983). Laminoid fenestrae are present in the clast as well, and these are common in subaerially exposed tidal flat laminite deposits (Tucker & Wright, 1990). It is possible that the fossiliferous calcilutite bed was laid down within a quiet nearshore lagoon or intertidal mudflat, which subsequently has been eroded away during contemporary Bathonian uplift and karstification of the landscape. The pebble was then transported to the hollow during flooding.

However, a second karstic origin for HQP2 and similar micritic clasts found at Hornsleasow is also proposed. Within covered and uncovered karsts, a superficial case-hardened micritic crust can be developed on the karstic surface of the limestone. These crusts faithfully trace the underlying topography of the karst and may be internally laminated. They have been termed laminated 'calcrete crusts' (Davies, 1991) and are composed of low magnesium calcite, although dolomitic 'dolocretes' can also form (Tucker & Wright, 1990). These crusts form either by *in situ* replacement of the original host limestone to micrite during pedogenesis (Kahle, 1977) or are precipitated as a type of case-hardening micrite cement upon lithified bedrock beneath a soil cover or upon exposed karst surfaces (Read, 1974). Those which develop without significant soil cover usually inhibit root and burrow penetration and tend to lack inclusions of detrital or

soil material. If the micritic clasts do represent the remains of a more pervasive crust, this could explain the strange absence of biotic activity within the host limestones underlying a thin *terra rossa* soil. Similar crusts are known to develop within the modern uncovered limestone scenery of Florida (Muller & Hoffmeister, 1968; Esteban and Klappa, 1983).

Davies (1991) described laminated calcrete crusts developed upon a Dinantian clay soil covered palaeokarstic surface in north Wales. The crust was partially dissolved and quite heavily brecciated by pedogenic penecontemporary processes, appearing only as eroded remnants upon upstanding hummocks on the surface after excavation (Davies, 1991). If an original micritic crust was present at Hornsleasow upon the karst, brecciation and weathering beneath the soil could have broken it up into the nodular horizon and weathered clasts seen upon the palaeokarst at the present. Spar-filled fenestrae have also been noticed within nodules derived from the micritic crust described by Davies (1991) and he suggests that they are brecciation fabrics formed by the desiccation associated with seasonal moisture fluctuation within the overlying clay soils. A similar hypothesis could be responsible for development of the rubbly layer or 'regolith' at the base of the clay paleosols at Hornsleasow. The cracks formed in the surface of the crust would become infilled with material from the overlying soils to develop the mixed carbonate-clay rubbly horizon and percolation of organic-rich soil waters would enhance formation of the *terra rossa*-like haematitic horizon. That a regolith layer is well-formed in the Hornsleasow soils, and exhibits a certain amount of mixing, suggests that the brecciation was ongoing during further clay paleosol maturation.

Apart from the evidence for the slight grain overpacking and pressure solution of some grains close to the dissolutional veins and vugs, it is highly unlikely that the limestone bed on which the karst is developed was deeply buried before uplift and weathering in the meteoric vadose zone in Bathonian times. In fact much of this apparent compaction probably occurred in the vadose weathering zone during karstification ('vadose compaction' of Tucker and Wright, 1990). There is little evidence of any post-karstic diagenetic changes made to the limestones below the palaeokarst surface at Hornsleasow, although further sparite cementation (as ferroan calcite) may have taken place in the reducing burial environment, but CDL studies would be needed in order to clarify this (section 3.6.1). It is probable that the calcite filled joints and cracks seen below the karst in the well-cemented calcarenites was precipitated into earlier cracks and joints during final burial.

4.4. Sequence of events during formation of the palaeokarst

The upper surface of the lower limestone is interpreted as a palaeokarst and the undulating topography an original feature of the karstic landscape. The limestones of the Chipping Norton Formation upon which the karst has developed are compositionally

bioclastic, oolitic and peloidal grainstones and were clearly deposited in a shallow subtidal palaeoenvironment (section 3.6.1). These limestones overly open marine oolite shoal sediments and the sedimentary sequence below the karst is interpreted as a regressive sequence culminating in subaerial weathering and soil formation. Interestingly, intertidal sediments (such as tidal flat laminites) appear to be absent in this sequence, although the appearence of several laminated calcilutite pebbles on the palaeokarstic surface may be the remnants of a former intertidal crust. A similar shallowing upwards sequence was described by Palmer (1979) from the White Limestone (Bathonian) of the English Midlands. The rarity of intertidal sediments in this sequence was interpreted by Palmer (1979) as indicating a low tidal range, and a humid climatic influence on sedimentation during the regression.

The steep sides of the clay-filled doline are reminiscent of the style of joint-block plucking which occurs commonly in coastal limestone areas, such as the Burren coast of western Ireland (Trudgill, 1985) and indicates a similar mode of formation for the Hornsleasow surface as it became emergent in early Bathonian times (Simms & Metcalf, in prep.). The eventual subaerial emergence of this irregular surface into the humid and subtropical climate prevalent during the Bathonian of the Cotswolds, resulted in the case-hardening of the karst.

Ivanovich & Ireland (1984) have suggested a figure of between ten and twenty thousand years for a one metre thick zone of case-hardened limestone to develop. Similarly, rinnenkarren of comparable depth with those at Hornsleasow were presumably initiated in the Caher River valley area of the Burren, County Clare, after the end of the last glaciation some 11 000 years ago and are therefore, this age (Trudgill, 1985). There has been considerable debate as to whether parts of the Burren formerly had a more extensive soil cover (for references, see Trudgill, 1985), beneath which many areas of rounded limestone pavement may have developed. However, the limestone pavement in the Caher River valley area is characterised by very large clints (some over $30m^2$), with wide flat surfaces and sharp edges (Simms & Metcalf, in prep.). They also show a suite of distinctive karren forms which indicate that the area never possessed any significant soil cover after the end of the last glacial (Simms & Metcalf, in prep.). Therefore, presuming that the main phase of case-hardening preceded the development of rinnenkarren at Hornsleasow, it seems likely that the karst surface was exposed to subacrial conditions for anything from about ten to twenty thousand years or more before any significant permanent soil cover developed (Simms & Metcalf, in prep.). However, the Burren karst developed in a post-glacial and temperate climate and cannot be directly comparable to the humid subtropical conditions in which the Hornsleasow karst formed. Relative rates of meteoric dissolution and diagenesis in the humid karsts of the Barbados and Bahaman area are extremely variable, and suggest exposure periods of between 10^3-10^5 years (Tucker and Wright, 1990).

However, despite the strong evidence that the surface of the Hornsleasow karst was indurated by an early reprecipitated low magnesium calcite cement and possible micritic crust, much of the exposed surface is now extremely soft and friable, so that it is only with careful excavation that its true morphology is revealed. Most of the rock samples gathered in 1993, and even those collected from unexcavated areas of the hollow by M.J. Simms in 1989, show significant decalcification and loss of cement (Figs. 4.6 & 4.7). Such a soft, porous texture is inconsistent with the prominent and sharp karst features, so it is clear that the surface must have undergone significant secondary dissolution, with the loss of much of the calcite cement developed during lithification.

The sharpness of the features exposed by the excavation are rapidly effaced by weathering and hence this phase of dissolution must have taken place at some time between the original deposition of the overlying paleosols and their ultimate removal by the excavating team in 1987-1989. Although this secondary dissolution may have occurred at any time during this interval, the presence of a considerable thickness of further Bathonian limestones above the unit throughout much of this time, would be an effective buffer for dissolution of the palaeokarstic surface by any initially acidic percolating water. Therefore, it seems most likely that much of the case-hardening cement was removed following the development of the overlying soil units by pedogenic brecciation and dissolution. The karstified limestones from underneath, and detrital fragments of oolite, within the paleosol units can be considered to be the regolith to the overlying soils (Davies, 1991; 'Ck' horizon of Foos, 1991). These have formed from more advanced subsoil alteration of the limestones and include features such as enhanced micritisation (Davies, 1991), rock brecciation, nodular secondary calcite or calcrete precipitation and insoluble residue enrichment (Davies, 1991).

The mineral soils are essentially made up of clays (Chapter 5), but do contain much limestone material in the form of rubble including pebble-, sand- and silt-sized clasts (section 5.6) and carbonate nodules (section 5.9). This suggests that overall the soils were essentially calcareous (i.e. by definition they contain more than 10% carbonate, Trudgill, 1985) and therefore, would themselves have acted as a buffer against acidic ground and rainwater percolating through them. However, the presence of much unoxidised plant material (section 8.3.3), with associated reduced iron in the form of iron pyrites (FeS_2; section 5.9.4) indicates that in parts the lower Grey Clay soil was constantly waterlogged and somewhat anoxic. The decay of the plant matter and the corpses of animals would have ensured a build-up of acidity in some parts of the soil. Living plants give off organic 'growth' acids such as humic and fulvic acid as products of metabolism (Trudgill, 1985). Upon decay of the plant material these growth acids will be released into the soil and will decompose to form simpler organic acids such as acetic, lactic, oxalic, citric and tartaric acids (Trudgill, 1985). Most of these acids are fairly weak, but concentrated upon a small area of limestone will leach calcium carbonate. The

reaction of these acids with the carbonates also functions to liberate carbon dioxide (CO_2), for example:

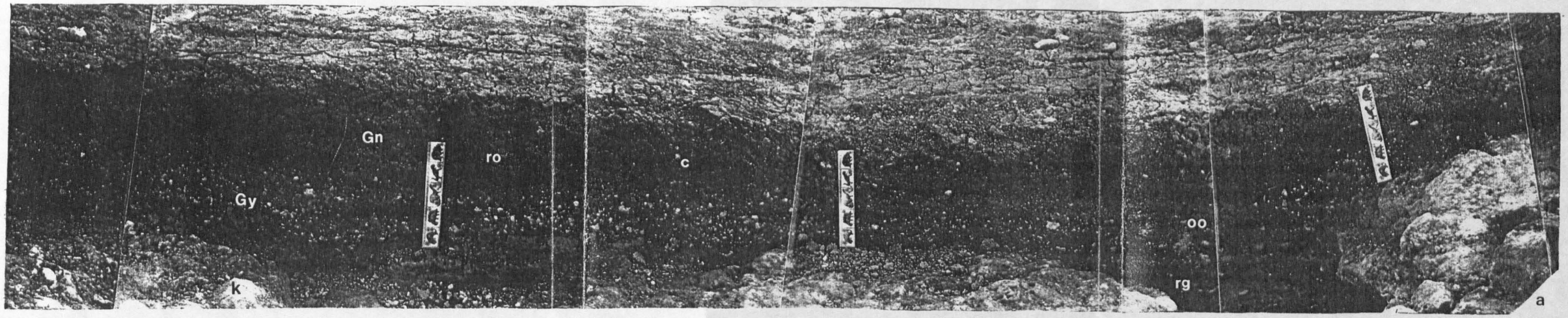
$$2HX + CaCO_3 \Rightarrow CaX_2 + CO_2 + H_2O \qquad (4.1)$$

where 'X' denotes the acid radical (Trudgill, 1985). The release of carbon dioxide into the soil is also an important factor in raising acidity levels in the porewaters, especially if the soil gas cannot escape. In this case much of the CO_2 will be present in the porewaters of the soil (4.2) and a proportion of this aqueous CO_2 combines with water to form carbonic acid (H_2CO_3) (4.3):

$$CO_{2\,(g)} + H_2O \Rightarrow CO_{2\,(aq)} + H_2O \qquad (4.2)$$

$$CO_{2\,(aq)} + H_2O \Rightarrow H_2CO_3 \qquad (4.3)$$

The impermeability of the waterlogged clay-rich paleosol unit at Hornsleasow would be conducive to localised acidic build-ups associated with the decomposition of plant and animal matter, leading to dissolution of the underlying karstic cements. The preservation of the calcite cement in regions beneath the large cetiosaur bones suggests that some sort of 'shadow effect' was involved, as the bones may have protected the underlying case-hardened limestone from dissolution and brecciation. The waterlogged conditions which prevailed during the formation of the lower paleosol horizon in the hollow indicate that the rinnenkarren and associated solution conduit no longer functioned effectively as a drain. This may reflect both a rise in the local water table and a blockage of the conduit associated with the deposition of the clays in the hollow (section 5.9). The implications of the karstic terrain and its association with the overlying clay paleosols are discussed in Chapter 6.



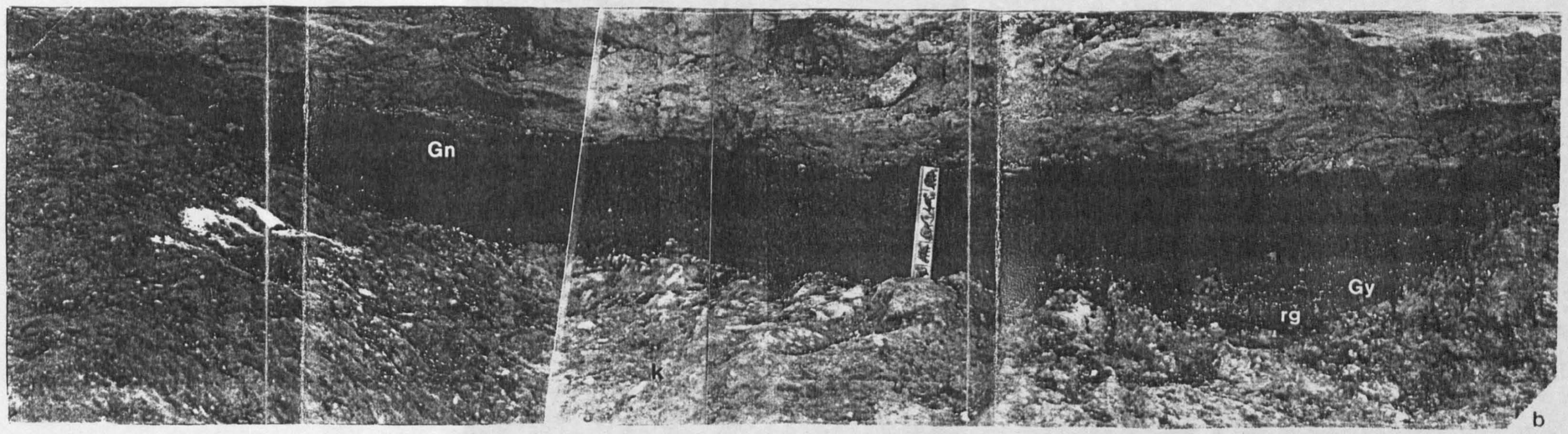


Figure 5.1. Panoramic shots of profiles across the clay deposits at Hornsleasow, showing (a) the section across the line B-B' on the excavation site A (Fig. 2.9) and (b) the section in the eastern wall of the unexcavated part of site B (Fig. 42.9). Abbreviations: 'k' karstic surface, 'Gn' Green Clay, 'Gy' Grey Clay, 'ro' rootlet, 'c' pedogenic carbonate clast, 'oo' oolite clast, 'rg' regolith. Scale = 0.3m, north (N) is indicated.

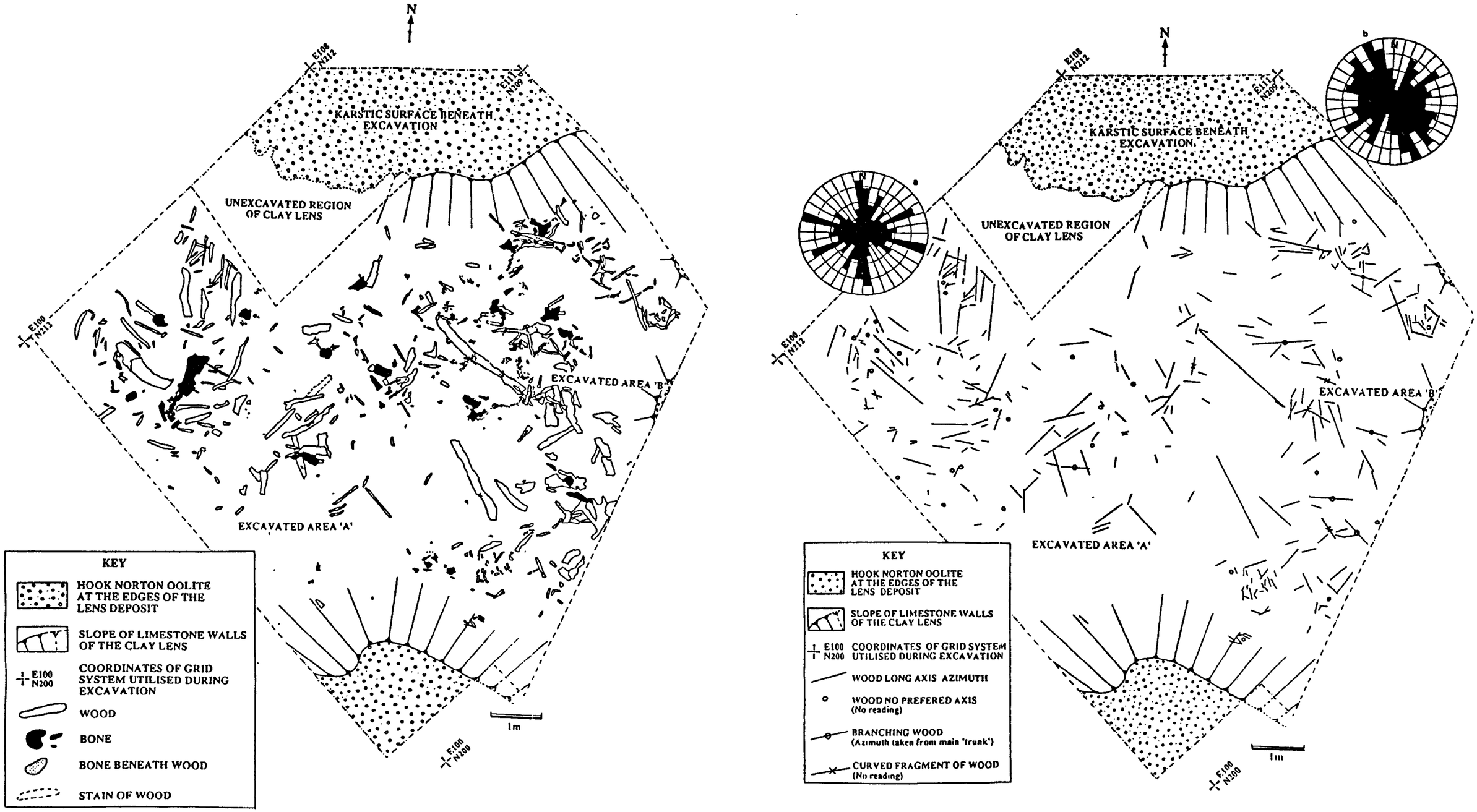


Figure 5.2. Sketch plans of the basal unit (karst and regolith layer) of the excavated site, showing distribution of wood and bone and the azimuth readings for the wood particles upon the gridded system (after Darlington, 1988, Metcalf *et al.*, 1992). Inset rose charts (a) and (b) show azimuth data for wood particles within the excavated areas 'A' and 'B' respectively.

accumulations many authors have depended upon preferred orientation of bones (e.g. Voorhies, 1969, Hunt, 1978, Shipman, 1981, Kreutzer, 1987) or wood particles (e.g. Gradzinski *et al.*, 1959) to illustrate alignment with a palaeoflow direction. To test for an alignment associated with palaeocurrent activity within the basal region of the lens, readings of the azimuth were measured along the long axis of the wood debris (for branching wood the reading was taken from the main 'trunk': Fig. 5.2) and some of the more elongate bones. Measurements were only taken for particles with a length:width ratio of over 3:1, as these were deemed to be more likely to show a preferred alignment to a palaeoflow direction (Potter & Pettijohn, 1977). It was possible to take a bearing with a Silver compass on 389 wood fragments and 170 elongated bones, and these data were used to construct rose diagrams (Fig. 5.3). As the material was seen to be confined to a single bedding plane (i.e. lying upon the karst) only azimuth data was read, this is because any measured angles of dip would be a reflection of the irregularities of the limestone pavement rather than true plunge (Fiorillo, 1988a). Therefore two-dimensional orientation rose charts were found to be appropriate for representing the data, rather than employing three-dimensional stereogram or equal-area plots (cf. Toots, 1965, Fiorillo, 1988a). The rose charts were constructed using the Rock-Edit™ and Rosy™ programs on the Apple Macintosh™, and show the data on 10° intervals between 0° (north) and 180°(south).

Almost exclusively, the bone and wood material from which measurements were taken, consisted of poorly preserved, mainly cylindrical fragments, lacking lateral projections, such as branches or roots in the case of the wood, and appendicular thickening, such as a root or epithycal mass in the case of the wood and bone material respectively. Hence, only azimuth trends rather than actual current directions were taken, as preferred upstream-downstream alignment could not be diagnosed (Fig. 5.4a, where the larger mass of the branches point downstream; Voorhies, 1969). The data was reflected over 360° to give a circular rose chart, which facilitated diagnosis of the palaeocurrent trends.

Fig. 5.3 shows the resulting rose charts for the wood material and bones within the basal portion of the clay lens. The data was also analysed statistically by the Rosy™ package to give a vector mean azimuth ($\underline{v}$) for the results (Curray, 1956) and the standard error (σ) on this calculated value, and these results are shown in Table 5.1. However, vector mean values are only equivalent to the palaeoflow direction rather than trends, and are only applicable to a unimodal distribution (Curray, 1956). Therefore, the Chi-Squared (χ^2) test was also applied to the raw data sets to test whether the distribution shown in the rose charts is truly random or whether the data exhibits any preferred alignment within the horizontal plane (Barford, 1985). The data were analysed between 0° and 180°, where 0° is equivalent to north, 90° is east and so on, in 10° intervals. The

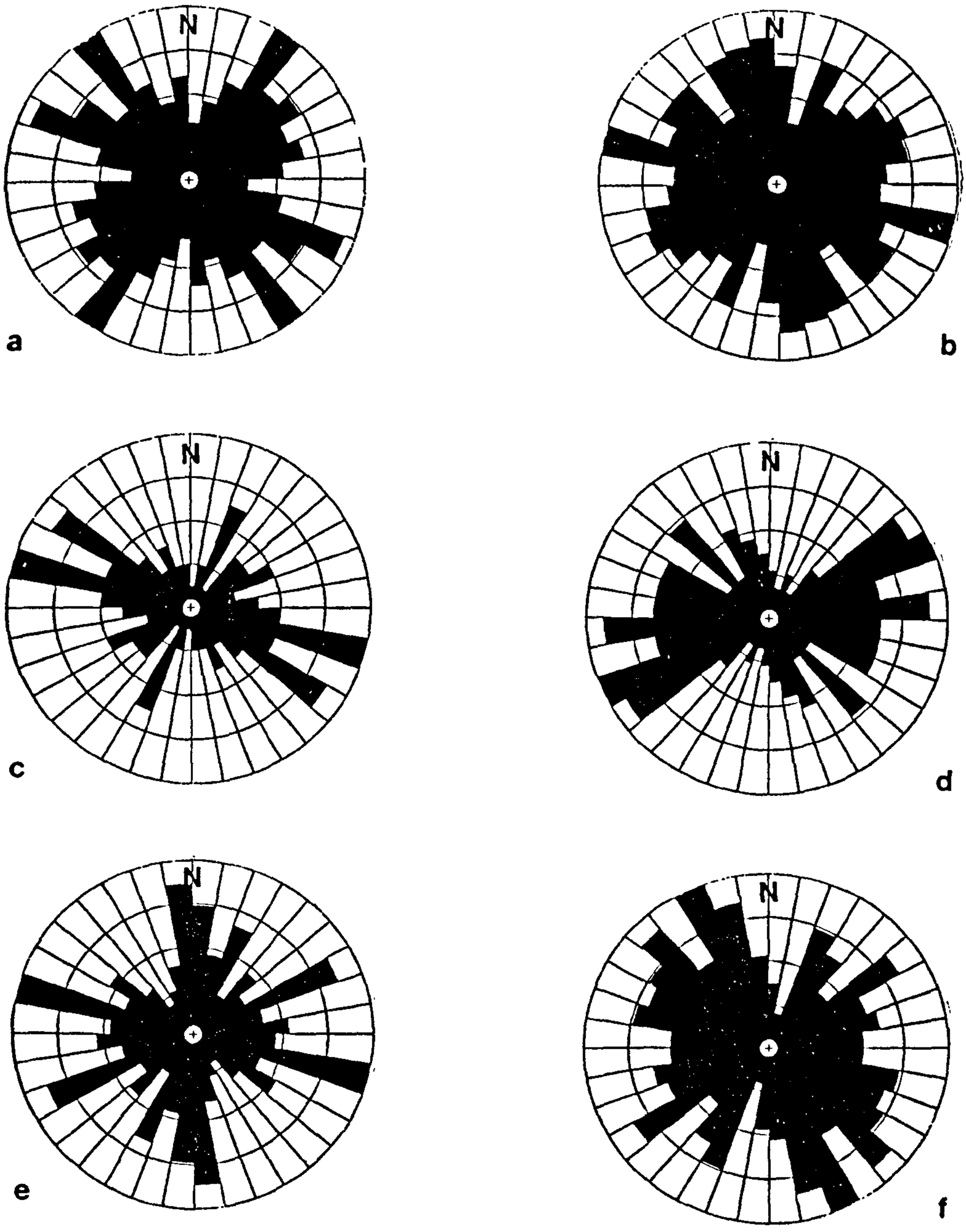


Figure 5.3. Rose diagrams of azimuth data taken from (a) 170 bone fragments, (b) 380 wood particles, (c) 59 isolated wood particles, (d) 102 wood particles with a L:W ratio of over seven, (e) 138 wood particles from excavated area 'A', and (f) 220 wood particles from excavated area 'B'. North (N) is indicated.

Data sets	Number of azimuth readings (N)	Vector mean ($\underline{v}$)	Standard error on $\underline{v}$ (σ)	Degrees of freedom (m)	Chi-squared number (χ^2)	Probability of being a random distribution (P)
Total bone	170	89°	153°	17	15.36	57%
Total wood	380	115°	55°	17	17.71	40%
Isolated wood	59	103°	23°	17	18.19	38%
Wood L/W > 7	102	86°	16°	17	21.75	20%
Site 'A' wood	138	56°	56°	17	26.09	6%
Site 'B' wood	220	129°	31°	17	18.28	37%

Table 5.1. Results of statistical analyses (vector mean, standard error and chi-squared tests) applied to the azimuth measurement data for the bone and wood resting on the palaeokarst surface, within the clay-filled hollow at Hornsleasow Quarry.

χ^2 test then works out how many azimuth readings in each data set are within the 10° intervals. If the distribution is assumed to be random, the χ^2 test predicts an average number of azimuth measurements which should be in each 10° interval, by dividing the number of measurements (N) by 18. Then the difference between the number of actual azimuth readings in each interval (n_i) and the predicted number of readings within the intervals (n_{pi}) gives an estimate of the probability of the distribution shown in the rose chart being random (Barford, 1985). This is expressed in terms of the χ^2 equation as:

$$\chi^2 = \sum_{i=1}^{m} \frac{(n_i - n_{pi})^2}{(n_{pi})} \qquad (5.1)$$

where m is the number of degrees of freedom, which in this situation is equal to the number of 10° intervals minus one, therefore m = 17 in all the analyses (Table 5.1). The results of the χ^2 test for each data set are shown in Table 5.1. From equation (5.1) it becomes clear that if n_i is equal to n_{pi} then χ^2 will be zero, and the distribution will be random. Therefore, values of χ^2 which approach zero represent data which is closer to a random distribution, whereas those with high χ^2 values suggest a degree of alignment. This degree of alignment can be expressed as a probability (p) of randomness and is found be checking the χ^2 results with a standard table of χ^2 values set against values of m (Barford, 1985). The probabilities are expressed for each rose diagram as a percentage and are shown in Table 5.1.

5.1.2. Results of the analyses

The rose chart for 170 bones (Fig. 5.3a) represents a moderately random distribution within a horizontal plane, (cf. Fiorillo, 1988a; Table 5.1) and there is no obvious alignment to a preferred palaeocurrent trend, although a vector mean of around east-west alignment with a huge associated error is suggested by Rosy™ (Table 5.1). However, this is not borne out by the rose-chart (Fig. 5.3a), which shows much stronger alignments in the northeast-southwest and northwest-southeast sectors, and it is apparent that these trends cancel each other in the dataset. However, the distribution pattern is almost certainly 'circular' and this is also suggested by the moderately high 57% probability of the data being random given by the χ^2 test (Table 5.1).

The second rose (Fig. 5.3b) shows the results of an analysis of a total of 380 wood fragments upon the base of the hollow (Fig. 5.2). The data and rose chart show alot of noise, and the χ^2 test reveals that there is a 40% chance of the distribution being random. The suggested vector mean (115°) indicates a trend through west-northwest to

east-southeast, with a correspondingly high standard error. In conclusion, although it is highly probable that the total wood data does indicate a palaeoflow trend, this has been damped by the noisy data. In order to try and eliminate some of the noise in the data set, experimental plots were also carried out upon sorted data from the total wood measurements (Figs. 5.3c-f & 5.2a,b).

The first analysis was based upon wood particles which were isolated from other accumulations of plant or bone debris on the palaeokarst surface. This helped to reduce the circular clustering and corresponding noisy azimuth data, which is shown by wood debris which have been caught in an eddy current or within a log jam (Fig. 5.4b). Also, because of the poor preservation of some of the wood material (as carbonised stains upon the limestone; section 8.3.2), it was not possible to differentiate between small associated 'sticks' and fragments of disarticulated 'branches' during the analysis. These two associations give strikingly dissimilar rose diagrams, as illustrated in Fig. 5.4a and could mask the original palaeocurrent trend. Therefore, discounting clustered wood material in the analysis would reduce repetition of azimuth readings or null readings taken from wood which were in life, lateral projections of larger branches. The number of particles within this data set is quite small (N = 59), whereas usually hundreds of specimens should be measured to ensure that any real alignment is not hidden by the inherent effects of noisy data (Jones & Dennison, 1970). Hence, the apparent alignment shown in Fig. 5.3c may in fact be chance rather than real. This is reflected by the χ^2 test yielding a 38% probability of the data being random. The suggested vector mean alignment trending west-northwest to east-southeast is quite obvious from the rose chart though (Fig. 5.3c) and is also close to that suggested for the total wood data (Fig. 5.3b) and could therefore, be a true palaeoflow trend. The error on this reading (23°) has been introduced by a large peak orientated perpendicular to the main to the maximum alignment (interpreted here as the palaeocurrent trend). The secondary peak can be explained by the movement of wood material within a water current which is both a factor of wood particle size, and water palaeodepth (Fig. 5.4c; Voorhies, 1969). Elongated wood or bone material which is entirely covered by water, tend to move along and come to rest in parallel orientations, whilst those which are slightly emergent are often orientated transverse to the current (Fig. 5.4c; Voorhies, 1969). This is complicated also by the method of particle movement within the water column, for instance dense or waterlogged bone material can sink rapidly and be carried along within the sediment load, whereas fresh bone floats within or at the surface of the water column (Fig. 5.4c; E. Cook, pers. comm., 1994). This is also true for wood which may remain afloat in favourable circumstances for several years (Haude, 1980, R.A. Spicer, pers. comm. to Simms, 1986). Therefore, it is expected that there should be an even distribution of azimuths parallel and transverse to palaeocurrent direction (Fig. 5.3c).

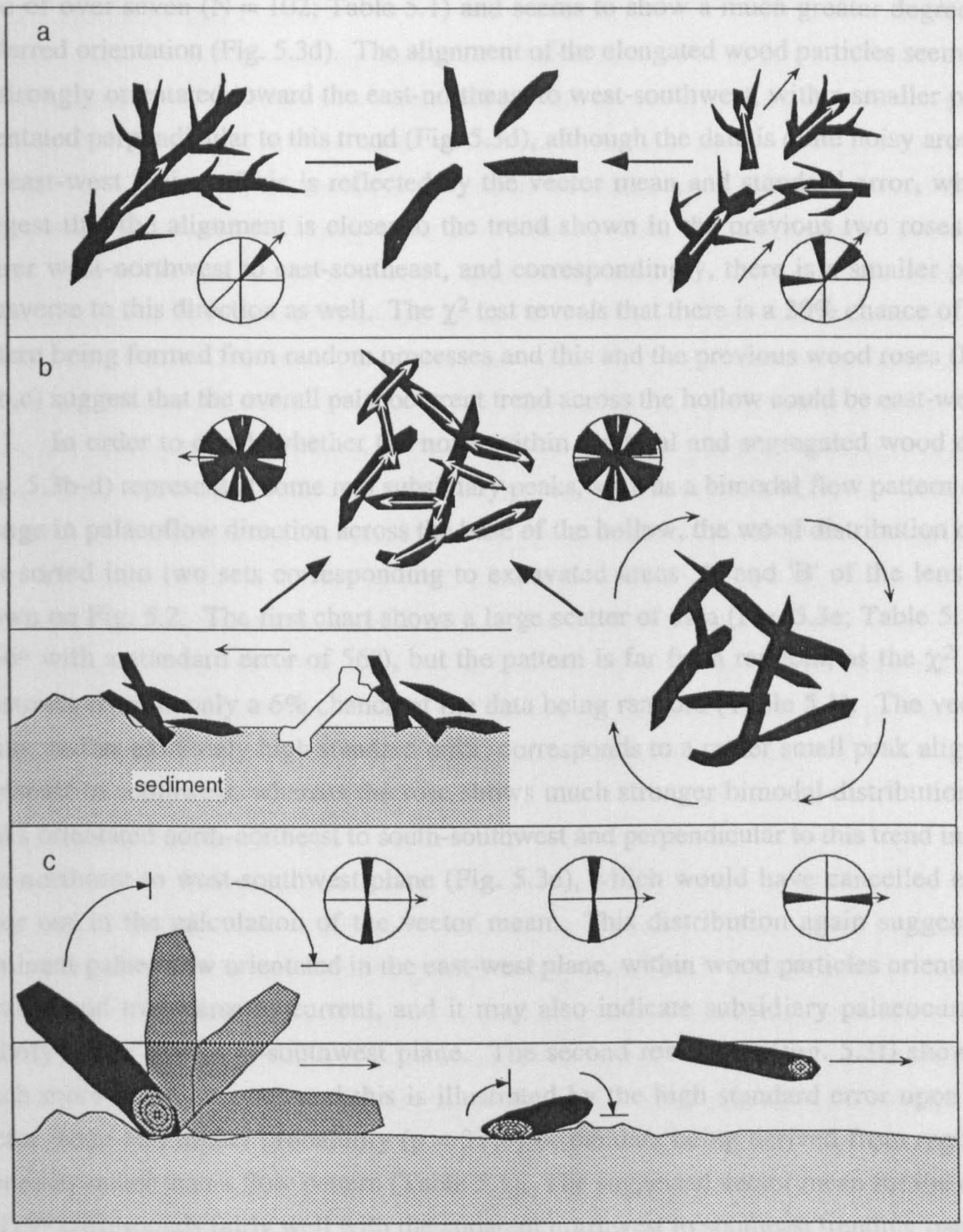


Figure 5.4. Cartoon showing potential hazards in ascertaining palaeoflow data from fossilised material: (a) Poor preservation of branched wood or several sticks; (b) Circular distribution of azimuth readings from a 'log-trap' or eddy-current; (c) Methods of motion for wood particles within a flowing water body.

The next plot corresponds to the most elongate wood particles, those with L:W ratio of over seven (N = 102; Table 5.1) and seems to show a much greater degree of preferred orientation (Fig. 5.3d). The alignment of the elongated wood particles seems to be strongly orientated toward the east-northeast to west-southwest, with a smaller peak orientated perpendicular to this trend (Fig. 5.3d), although the data is quite noisy around the east-west region. This is reflected by the vector mean and standard error, which suggest that the alignment is closer to the trend shown in the previous two roses, of nearer west-northwest to east-southeast, and correspondingly, there is a smaller peak transverse to this direction as well. The χ^2 test reveals that there is a 20% chance of the pattern being formed from random processes and this and the previous wood roses (Fig. 5.3b,c) suggest that the overall palaeocurrent trend across the hollow could be east-west.

In order to check whether the noise within the total and segregated wood data (Fig. 5.3b-d) represented some real subsidiary peaks, such as a bimodal flow pattern or a change in palaeoflow direction across the base of the hollow, the wood distribution data was sorted into two sets corresponding to excavated areas 'A' and 'B' of the lens, as shown on Fig. 5.2. The first chart shows a large scatter of data (Fig. 5.3e; Table 5.1, $\underline{v}$ = 56° with a standard error of 56°), but the pattern is far from random, as the χ^2 test illustrates there is only a 6% chance of the data being random (Table 5.1). The vector mean, and its extremely high standard error, corresponds to a rather small peak aligned northeast to southwest, whereas the rose shows much stronger bimodal distribution of peaks orientated north-northeast to south-southwest and perpendicular to this trend in the east-northeast to west-southwest plane (Fig. 5.3e), which would have cancelled each other out in the calculation of the vector mean. This distribution again suggests a dominant palaeoflow orientated in the east-west plane, within wood particles orientated parallel and transverse to current, and it may also indicate subsidiary palaeocurrent activity in the northeast-southwest plane. The second rose chart (Fig. 5.3f) shows a much more noisy data set, and this is illustrated by the high standard error upon the vector mean and higher probability (p = 37%) of the data being derived from random processes rather than a flow pattern (Table 5.1). The suggested vector mean for the data of 129° corresponds fairly well with the apparent northwest to southeast trending pattern illustrated upon the rose chart (Fig. 5.3f) and there is a slight indication of a subsidiary peak orientated perpendicular to this preferred trend. The higher probability of the distribution being random in Area B, can be explained by the higher proportion of circularly orientated wood material in this region (Figs. 5.2 & 5.3) associated with log jams and ?eddy currents (Fig. 5.4b).

5.1.3. Palaeoflow conclusions

In conclusion, there is then a suggestion of a weak palaeoflow within this basal layer, which suggests that the large wood material was not growing within the hollow. It was more probably derived from local vegetated topographical highs on the surrounding floodplain. The low-lying karstic hollow and the cetiosaur bones contained therein may have acted as a trap for branching wood debris during times of flooding upon the karstic plain. The wood and bones may have been transported to the present site during a temporary flood upon the karstic landscape or the site may represent the distal section of an abandoned ?east-west trending stream later modified by soil-forming processes - the former hypothesis is favoured because of the lack of a strong palaeoflow direction and because of the palynofloral evidence (section 8.3.1). Flooding events would also deposit terrigenous mud, vertebrate cadavers and isolated bone material derived from the surrounding floodplain into the hollow. The general palaeoflow direction seems have been orientated either east-west or north-northwest to south-southeast, and therefore, flooding could have been derived from either the northern Midlands cratonic area or from the Oxfordshire shallows in the east (section 6.2). The deposition of more sediments and formation of a boggy bioturbated soil within the hollow may have altered the position and attitudes of the bones and wood, before stabilisation during final burial.

5.2. Paleosols - an introduction

A fossil soil or paleosol is defined as the remains of an ancient soil buried by later sediments (Retallack, 1990), and paleopedology is the study of ancient soils including relic soils still at the surface and being modified by modern day soil-forming processes, as well as the true fossil soils which occur within ancient sedimentary sequences. The overall aim of paleopedology is to reconstruct terrestrial palaeoenvironments, as the identification of paleosols within a sequence are the most direct evidence for indicating subaerial exposure and weathering (Wright, 1986). In studying ancient soil profiles, various geological processes can be elucidated upon, for instance the features of a paleosol can be related to the atmospheric conditions, vegetation types and weathering processes prevalent at the time of soil formation (Retallack, 1976, 1977,1986). In this capacity, paleosols can be considered to be the "trace fossils of past ecosystems and as the preservational environment for many kinds of fossils" (Retallack, 1990, p. 7).

5.2.1. General features of paleosols

The clay deposits within the fossiliferous lens discovered at Hornsleasow Quarry is considered by this author to be the remains of a paleosol profile developing upon the karstified surface of the Chipping Norton Formation limestones in near-coastal terrestrial

conditions (Metcalf *et al.*, 1992, Metcalf, 1993). In the next sections, the rationale behind the identification of the clay lens as a fossil soil profile is set out and explained. In doing so it is necessary first to explain the main features of a typical modern soil profile and then to extrapolate these features into the fossil record.

A soil is defined as the material forming at the surface within the 'zone of interaction' between the atmosphere and bedrock of a terrestrial planet (Retallack, 1990). This material is altered *in situ* from the parent rock by the combined effects of physical, chemical and biological weathering processes (Retallack, 1990). A typical soil profile consists of three layers (Fig. 5.5a), which are:

(1) *The solum.* This is the layer at the surface which is most altered by soil-forming processes or 'pedogenesis'. It is often pervaded by abundant root traces and is generally quite different from the original parent material (Retallack, 1990).

(2) *The saprolite.* This is the zone of mixing and consists of a mixture of sediment or bedrock and the altered products of pedogenesis (i.e. solum). This layer may show relict features of the underlying parent material, such as ghost bedding and may also be pervaded by root traces (Retallack, 1990).

(3) *The bedrock.* This is the underlying, unaltered parent material. It is commonly called the 'regolith' or the 'country rock' (Retallack, 1990).

Soils form from a combination of mechanical, biological and chemical weathering of the original regolith, they continue to form until they are finally buried by a cover of sediment or are removed through erosion. The events which lead up to soil formation, are fairly well known and begin with the original rock being brought to the surface. This dry rock surface may be cracked and jointed, allowing colonisation by small plants and burrowing organisms. If the climate is humid and warm, this initial colonisation will be followed by shrubs and eventually woodland growth. As the rock is burrowed and penetrated by roots, and altered by the chemical and mechanical weathering processes, the original features of the regolith become destroyed and a saprolitic zone develops. Remnant features are present in very weakly developed soils and should be looked for in studying the soil profile.

Most fossil soil profiles will show a variation upon that illustrated in Fig. 5.5a, although paleosols are not usually as loose, cracked and crumbly as soils presently forming at the surface. Most paleosols can appear featureless and fairly massive in comparison to the surrounding country rock and this could probably explain the lack of investigation by Channon (1950) of the previous clay lens at Hornsleasow Quarry. Paleosols are usually compacted and may be altered by burial diagenesis (section 5.11), but it is possible to recognise original features in most fossil soils (Retallack, 1990). The three most useful and easily identifiable large-scale features of fossil and modern soils are root traces, soil horizons and the soil structure (Retallack, 1990).

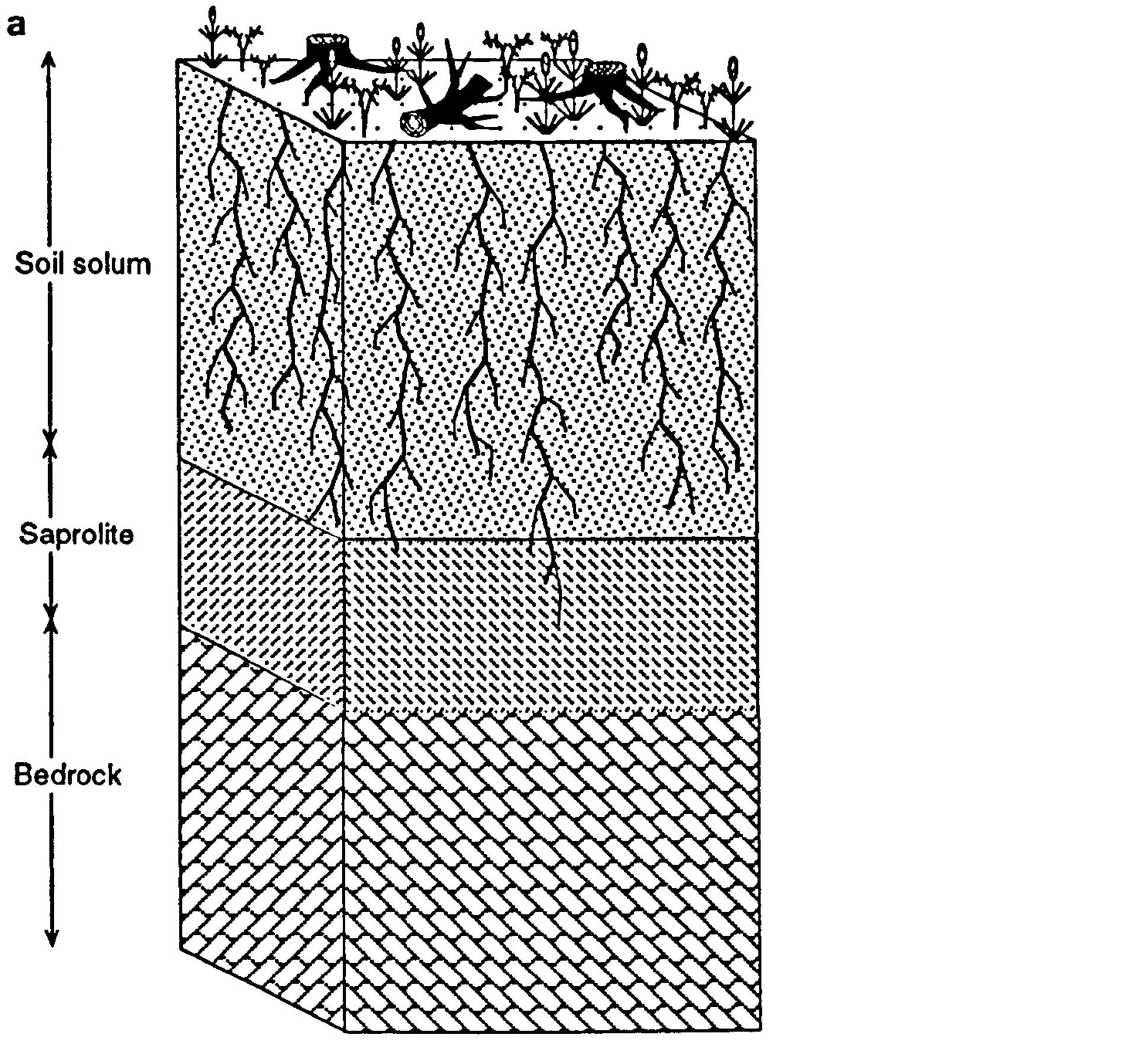


Figure 5.5. Typical weathering profile (a) showing main features of a soil. (b) shows the large-scale and microscopic structures seen within the main saprolitic horizon of the soil profile (after Retallack, 1990).

In order to study these features and the petrology of the Hornsleasow clay soils in the laboratory, four 0.5x0.5m square blocks were removed from the lens during excavation. These where cut from one of the diagonal unexcavated 'walls' of clay (Line B"-B"' on site plan, Fig. 4.2) in excavation Site B (Fig. 5.6). The blocks were numbered S9-S12, and were stabilised from above by applying a styrofoam case to the upper surface. Once the case had hardened the blocks were cut away from the wall using complex manual digging implements or 'spades' (Darlington, 1989). The block sections were then wrapped in aluminium foil and cling film, before being encased in tough, impermeable styrofoam and sent back to the laboratory for examination. This ensured that the original colour, moisture content and redox conditions of the soils could be studied in the laboratory. From these sections soil profiles were made (Fig. 5.7) and samples taken for chemical, mineralogical and petrographic analyses, in order to elucidate small-scale features such as soil structure, grain composition and the identification of soil nodules. These analyses also helped to provide an indication of the weathering profile and time-span of soil formation (section 5.10), and any diagenetic changes which may have occurred after the soil was buried (section 5.11). Finally, the Hornsleasow soil profile was tentatively classified and assigned a soil type based upon modern schemes for soil taxonomy (section 5.12). The main aim of this and the previous chapter is to provide clues as to the palaeoenvironmental conditions prevalent in the early Bathonian. The implications of the soils and underlying palaeokarst at Hornsleasow are discussed in the next chapter.

5.3. Soil horizon description and soil profiles

The general soil profile shown in Fig. 5.5a, can be further subdivided into many different categories or soil horizons, defined upon the basis of the materials which comprise them. These can depend upon the original parent rock, the conditions under which the soils were formed and can also be modified by burial. It is useful in the identification and study of fossil soil profiles to identify individual soil horizons, as this aids identification of the overall soil type and facilitates studies of small and large scale features.

Soil horizons are characterised and tagged by special shorthand labels, some of which are shown in Table 5.2 (after Retallack, 1990, taken from the *1982 U.S. Soil Survey Manual,* Guthrie & Witty, 1982). These include master horizons , such as the solum ('O', 'A', 'E' horizons), saprolite ('B', 'C' horizons) and bedrock ('R' horizon), and subordinate descriptions which appear as superfixes after the initial description (Table 5.2). Gradation between master horizons can also occur and are expressed in terms of both shorthand labels, but with the dominant horizon presented first, for example 'BC' indicates a combination of 'B' and 'C' characteristics but with the former horizon dominant (Retallack, 1990).. A combination of two letters with a diagonal slash

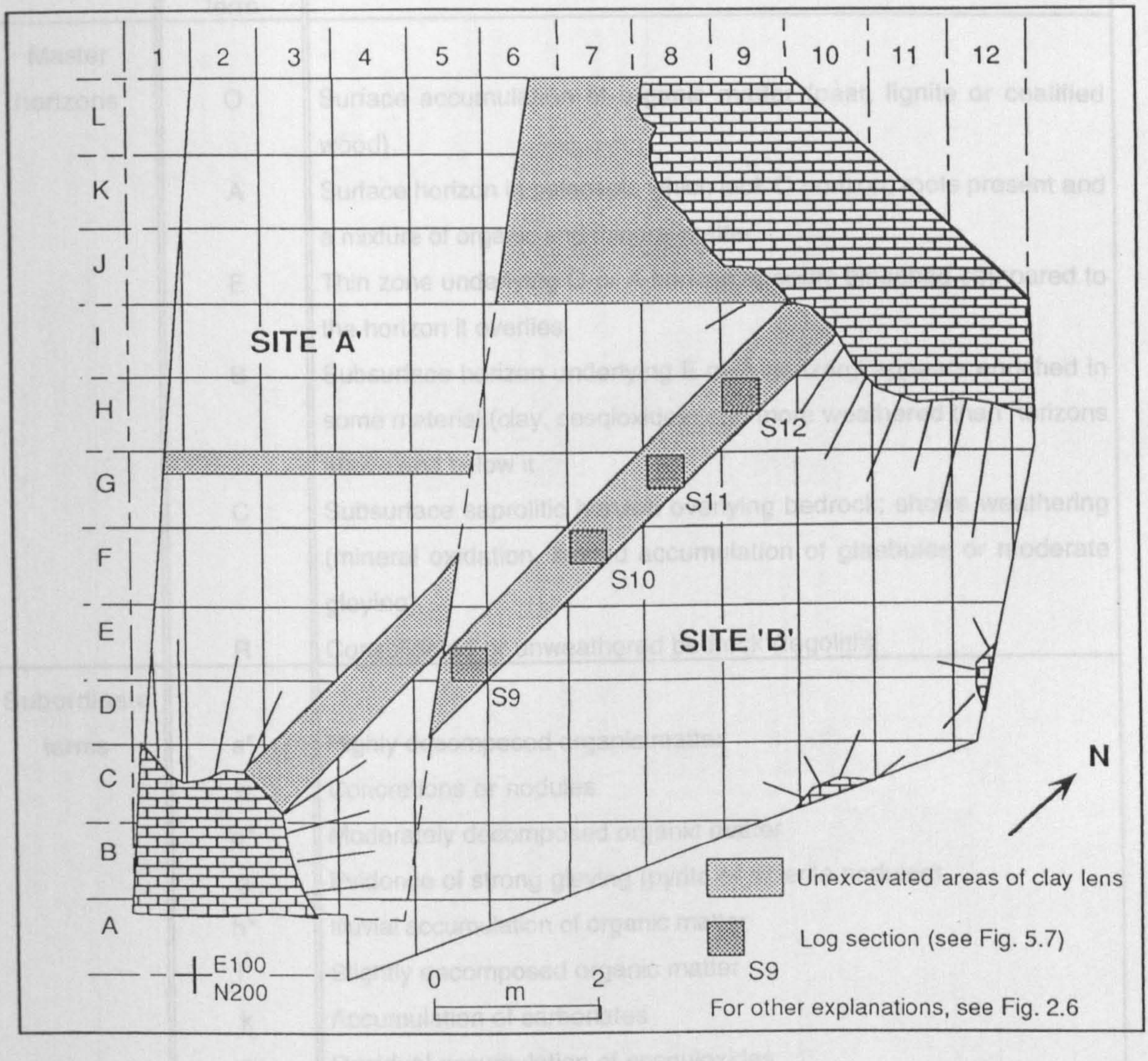


Figure 5.6. Positions of the soil blocks (S9-S12) removed from the unexcavated areas of the clay lens in excavated site 'B'. North (N) is indicated.

Category	Shorthand term	Description
Master horizons	O	Surface accumulation of organic matter (peat, lignite or coalified wood)
	A	Surface horizon in paleosols which lack O horizon; roots present and a mixture of organic and mineral matter
	E	Thin zone underlying O or A horizon;appears bleached compared to the horizon it overlies
	B	Subsurface horizon underlying E or A horizon;I appears enriched in some material (clay, sesqioxides) and more weathered than horizons above and below it
	C	Subsurface saprolitic horizon overlying bedrock; shows weathering (mineral oxidation, limited accumulation of glaebules or moderate gleying)
	R	Consolidated or unweathered bedrock (regolith)
Subordinate terms	a*	Highly decomposed organic matter
	c	Concretions or nodules
	e*	Moderately decomposed organic matter
	g	Evidence of strong gleying (pyrite or siderite nodules)
	h*	Illuvial accumulation of organic matter
	i*	Slightly decomposed organic matter
	k	Accumulation of carbonates
	o	Residual accumulation of sesquioxides
	r*	Weathered or soft bedrock
	s*	Illuvial accumulation of sesquioxides
	t*	Accumulation of clay
	w*	Coloured or structured B horizon

Table 5.2. Standard U.S. Soil Survey descriptive shorthand for labelling paleosol horizons (after Retallack, 1990, Guthrie & Witty, 1982). Note that some of the subordinate terms are more important than the others, these are denoted by an asterisk and should be quoted first in all combinations with other letters and not used in combinations with each other.

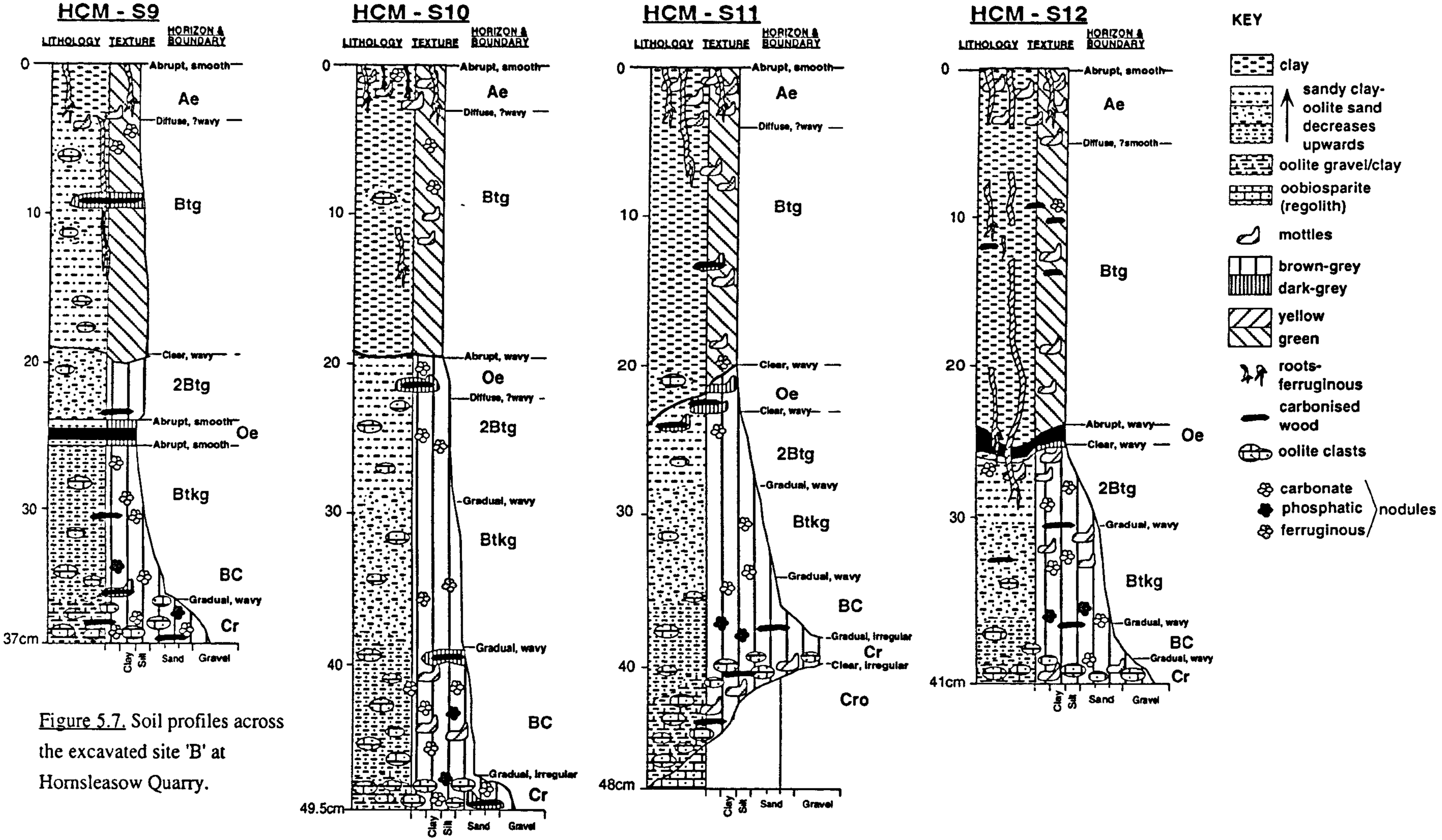


Figure 5.7. Soil profiles across the excavated site 'B' at Hornsleasow Quarry.

between them (e.g. 'E/B') indicates that the 'B' horizon is dominant with tongues or inclusions of material like an 'E' horizon within it (Retallack, 1990).

The determination and characterisation of the horizons within the Hornsleasow soil profile was completed from a field and from a laboratory study of the four blocks, which are figured in Fig. 5.7. The soil profiles are fairly uniform across the site and consist of a strong bipartite subdivision (Figs. 5.1 and 5.8), which were termed in the field as the lower 'Grey Clay' and upper 'Green Clay' (section 2.4.1; Vaughan, 1988, 1989). These two main horizons were originally subdivided further by a complex system of context numbers given to local variants by the excavators (Fig. 4.2; Darlington, 1988). For instance, the Grey Clay was given two context numbers, depending upon where it was bagged upon the site: 1007 for site A and 1022 for site B; and the Green Clay has no less than eight different context numbers reflecting local colour variations and collector preferences (Darlington, 1988). In order to simplify these results, I choose to characterise the horizons within the soil profiles based only upon the bipartite subdivision of Vaughan (1988) and the soil science subdivisions given in Table 5.2.

In all the following descriptions the Green and Grey Clay units shall be considered to be two separate soil profiles. The *terra rossa* clays directly overlying the palaeokarst and the karst surface itself is considered to be a separate soil unit (section 4.4.), although pedogenic effects and bioturbation within the overlying Grey Clay have caused overprinting of the *terra rossa*. The clays which comprise the two units are considered to be allochthonous (section 5.10), although some accumulation of soil matrix from the underlying limestone is also indicated and much mixing has occurred. In order to characterise the paleosol profiles, soil horizon thickness and identification, grain size, colour and the reaction with acid were all ascertained from the large blocks, and the results summarised in Fig. 5.7. These are preliminary 'field' descriptions, and their primary use was to provide a guide for further petrographic, chemical and mineralogical investigations of smaller scale soil structures (sections 5.5 - 5.9).

5.3.1. Horizon thickness, boundary description and interpretation

The soil profiles shown in Fig. 5.7 are fairly uniform across the clay lens, they illustrate a distinct bipartite subdivision between the upper Green Clay and lower Grey Clay (Fig. 5.8), and a much more diffuse boundary between the Grey Clay and the underlying limestones (Fig. 5.1). This is typical of many fossil soils occurring in sedimentary sequences (Retallack, 1990), and which show a sharply truncated top surface and gradational lower boundary between the saprolite and country rock. The abrupt and smooth surface at the top of the Green Clay (Fig. 5.1) is an erosional contact and can be considered to be the palaeo-land surface. Upon excavation the surface was found to be suncracked and fragments of weathered bone and oxidised traces of wood debris

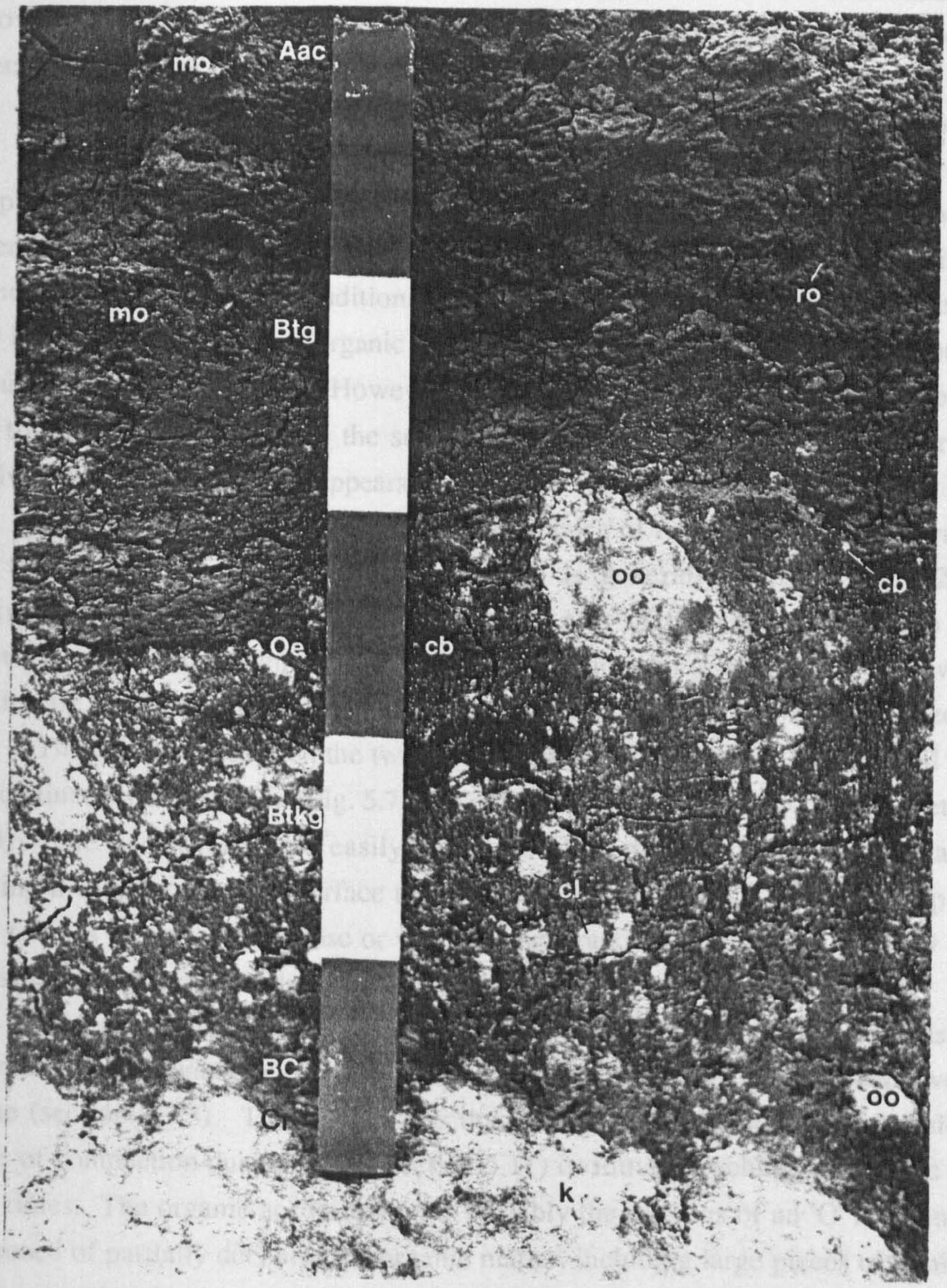


Figure 5.8. The Grey Clay and Green Clay soil profile with underlying karst surface in the region of soil block S9 (excavated area 'B'). Abbreviations: 'cb' carbonaceous matter; 'cl' calcareous glaebules; 'k' palaeokarstic surface; 'mo' mottling; 'oo' oolitic limestone clast; 'ro' rootlet.

discovered (Fig. 5.1; Metcalf *et al.*, 1992). The top surface was apparently covered by vegetation and not permanently waterlogged.

The Green Clay soil profile is characterised by only two distinct horizons (Fig. 5.7). It could be considered to be the top part of the underlying Grey Clay soil profile or a separate paleosol unit. I prefer the latter theory, as there is strong sedimentary, mineralogical and palaeontological evidence that the soils developed in slightly different, but not mutually exclusive, conditions (section 5.10). The surface horizon of the Green Clay lacks an accumulation of organic matter and is basically similar in composition to the subsurface clay-rich horizon. However, it is distinctly mottled and contains abundant root traces which extend from the suncracked surface (Fig. 5.1). This 'A' horizon is relatively thin (30-50mm) and appears to have been well oxidised.

The degree of colour mottling and abundance of rootlets decreases downwards into the underlying clay-rich horizon, which is generally between 0.1-0.25m in thickness. If this horizon is considered to be part of the same soil profile as the underlying Grey Clay, then it is best termed a 'Bt' horizon (Fig. 5.7). If however, the profiles are unrelated it is probably better characterised as a 'Ct' or even 'Cr' horizon.

The boundary between the two clays is quite well defined in the four soil blocks by a distinct colour variation (Fig. 5.7) and can be described as clear to abrupt (Retallack, 1990). The two layers were easily separated along this contact during excavation forming a continuous upper surface to the Grey Clay (Vaughan, 1989). The top of the Grey Clay is somewhat undulose or wavy in appearance (Fig. 5.8) and this may reflect clay swelling under seasonal environmental conditions.

Underlying this surface in some parts of the clay lens are the remains of an accumulation of organic matter, which now consists of drab-haloed coalified wood or lignite (section 8.3.3). The layer is less than 30mm in thickness, but this could be a factor of compaction during burial (section 5.11) or illuvial leaching, during the Green Clay times. The organic accumulation is possibly the remains of an 'O' horizon. It is composed of partially decomposed organic matter, including large pieces of lignite and can be considered to be an 'Oe' horizon. The thin layer of drab pale grey coloured clay underlying the organic horizon, is interpreted as an 'E' horizon (Table 5.2). However, illuvial accumulation of the organic matter by percolating waters, and bioturbation or during sedimentation cannot be ruled out.

The soil profile of the Grey Clay is generally made up of successive layers of less calcareous waterlogged or 'gleyed' brown-grey clay. In the main, the calcareous material has been derived from the underlying limestones and the soil is integrally related to the palaeokarstic surface (sections 4.4 & 4.5). The profile is made up of a succession of thin 'B' horizons composed of essentially gleyed clay (hence the suffix 'tg') and the residual carbonate ('k') and is relatively thin at between 0.1-0.2m in comparison to the more

massive Green Clay unit (Fig. 5.7). The different 'B' horizons are somewhat arbitrary and grade over several centimetres into one another, and can be considered to have gradual or diffuse boundaries (Retallack, 1990).

The 'B' horizon gradually becomes more oolitic toward the base of the profile and gives way to a mixed clay-carbonate sand deposit ('BC'), which is full of oolitic clasts (Fig. 5.8). This horizon also contains much lignified wood and pedogenic glaebules (section 5.9). The 'BC' horizon is no more than 30-50mm in thickness and gradually gives way to a much more lime-rich saprolitic unit, the 'C' horizon (Figs. 5.7 & 5.8). This regolith horizon is characterised by the abundance of ferruginous carbonate sand, clay clasts, organic matter and soft weathered pebbles of oolitic bedrock. It also contains the partial remains of the *Cetiosaurus* dinosaur and indurated iron-stained karstic pebbles (section 4.5). This horizon directly overlies the friable haematitic surface of the limestone, which can also be considered to be a 'C' horizon, as it has undergone much chemical alteration and mineral oxidation (section 4.4.2).

5.3.2. Colour

The colour of the soil units was described and distinguished using a standard *Munsell^R Color Chart* and also given a reference Munsell^R value (Rock-Color Chart Committee, 1991). Munsell^R values describe primary colours or 'hues' and take the form of 5R or 10YR. This is followed by a number for the 'value' of the degree of lightness of the primary hue, which is expressed on a scale from 8 - 2. Lastly a number is given for the 'chroma', which is the degree of saturation of colour from neutral grey to purple (Rock-Color Chart Committee, 1991). For example a pale greyish orange-pink is described as 10R followed by 8 and then 2, so it's Munsell^R description is 10R 8/2.

It is important to establish the colour of wet sediments or soils in the field, as well as from the soil blocks. This is because as the soils dry out, they become paler, and although desiccation was inhibited in storage, moisture loss was still apparent. Excavation and laboratory storage may also induce originally drab grey or green soils to become warmer in hue, this is because the ferrous iron within them becomes oxidised upon exposure to air, to yellowish ferric oxyhydrites (Retallack, 1990). The results of this determination are shown in Table 5.3.

The colour determination was carried out as a preliminary step towards palaeoenvironmental assessment of the soil horizons and as a check upon weathering and diagenetic change. For instance, the yellow mottled effect is typical of an aerobic 'A' horizon associated with the churning effects of root activity. Bright green colours (Fig. 5.8) are not usually associated with modern soils, although they are common in paleosols (V.P. Wright pers. comm., 1993) and this might suggest that the surrounding clays were slightly reducing or seasonally gleyed. The green coloration might also be due to

SOIL UNIT	HORIZON	FEATURE	MUNSELL[R] VALUE	DESCRIPTION	REACTION WITH ACID	CARBONATE CONTENT
S9	'Aac'	matrix	5GY 5/2	dusky yellow green	unreactive	noncalcareous
S9	'Aac'	mottles	5Y 5/6	light olive brown	unreactive	noncalcareous
S9	'Btg'	matrix	10Y6/2	pale olive	slight	very weakly calcareous
S9	'2Btg'	matrix	5Y 4/1	olive grey	slight	very weakly calcareous
S9	(?'O')	matrix	5G 2/1	greenish black	unreactive	noncalcareous
S9	'Btkg'	matrix	5YR 4/1	brownish grey	reactive	calcareous
S9	'BC'	matrix	5YR 4/1	brownish grey	reactive	calcareous
S9	'Cr'	matrix	5YR 6/1	light brownish grey	reactive	calcareous
S9	'Cr'	micrite glaebules	5Y 8/4	greyish yellow	vigorous	strongly calcareous

Table 5.3. Description of the colour and reaction with dilute (10%) hydrochloric acid for horizons and features within the Hornsleasow paleosol profile.

SOIL UNIT	HORIZON	FEATURE	MUNSELL[R] VALUE	DESCRIPTION	REACTION WITH ACID	CARBONATE CONTENT
S10	'Aac'	matrix	5GY 5/2	dusky yellow green	unreactive	noncalcareous
S10	'Aac'	mottles	5Y 5/6	light olive brown	unreactive	noncalcareous
S10	'Btg'	matrix	5GY 5/2	dusky yellow green	unreactive	noncalcareous
S10	'Btg'	micrite glaebules	5Y 8/4	greyish yellow	vigorous	strongly calcareous
S10	'2Btg'	matrix	10YR 4/2	dark yellowish brown	slight	very weakly calcareous
S10	'Btkg'	matrix	10YR 4/2	dark yellowish brown	reactive	calcareous
S10	'Btkg'	iron nodule	10YR 6/6	dark yellowish orange	slight	very weakly calcareous
S10	'BC'	matrix	5YR 4/1	brownish grey	slight	very weakly calcareous
S10	'BC'	oolitic clast	10YR 8/6	pale yellowish orange	vigorous	strongly calcareous
S10	'Cr'	matrix	5YR 5/2	pale brown	vigorous	strongly calcareous

Table 5.3. continued....

SOIL UNIT	HORIZON	FEATURE	MUNSELL[R] VALUE	DESCRIPTION	REACTION WITH ACID	CARBONATE CONTENT
S11	'Aac'	matrix	10Y 6/6	dark greenish yellow	unreactive	noncalcareous
S11	'Aac'	mottles	10YR 6/6	dark yellowish orange	unreactive	noncalcareous
S11	'Btg'	matrix	10Y 6/2	pale olive	slight	very weakly calcareous
S11	'Oe'	matrix	N2	greyish black	unreactive	noncalcareous
S11	'E/B'	matrix/mottles	N5	medium grey	slight	very weakly calcareous
S11	'2Btg'	matrix	10YR 4/2	dark yellowish brown	slight	very weakly calcareous
S11	'Btkg'	matrix	5YR 4/1	brownish grey	reactive	calcareous
S11	'BC'	matrix	10YR 5/4	moderate yellowish brown	vigorous	strongly calcareous
S11	'Cr'	matrix	10YR 5/4	moderate yellowish brown	vigorous	strongly calcareous
S11	'Cro	iron-stained mottles	10YR 6/6	dark yellowish orange	slight	very weakly calcareous

Table 5.3. continued....

SOIL UNIT	HORIZON	FEATURE	MUNSELL[R] VALUE	DESCRIPTION	REACTION WITH ACID	CARBONATE CONTENT
S12	'Aac'	matrix	5Y 6/4	dusky yellow	unreactive	noncalcareous
S12	'Aac'	mottles	10YR 6/6	dark yellowish orange	unreactive	noncalcareous
S12	'Btg'	matrix	5Y 5/2	light olive grey	unreactive	noncalcareous
S12	'Oe'	matrix	N2	greyish black	unreactive	noncalcareous
S12	'E/B'	matrix/mottles	10YR 4/2	dark yellowish brown	slight	very weakly calcareous
S12	'2Btg'	matrix	5Y 5/2	pale brown	slight	very weakly calcareous
S12	'Btkg'	matrix	5YR 4/1	brownish grey	reactive	calcareous
S12	'BC'	matrix	5Y 4/1	olive grey	reactive	calcareous
S12	'Cr'	matrix	10YR 8/2	very pale orange	vigorous	strongly calcareous

Table 5.3. continued....

diagenetic changes, such as alteration of the clay minerals and reduction of ferric iron by marine water percolation (V.P. Wright pers. comm., 1993, Davies, 1991) or reduction of ferric iron during burial gley (Retallack, 1990) (section 5.11).

The grey-brown colour of the underlying Grey Clay (Fig. 5.8) is common amongst soils which are permanently waterlogged or gleyed. The colour becomes even more grey near organic matter and can be considered to form 'drab-haloes' (Retallack, 1990). The various theories concerning the formation of organic matter drab haloes are considered in detail in section 5.9.4. However, the grey coloration may be derived from a genuine reduction of iron around the organic matter or might be stained by fine specks of organic matter dispersed throughout the clay matrix. The various colours of the underlying karstified limestones have been described in a previous chapter (section 4.5) and shall not be referred to here.

5.3.3. Grain size analysis

Information upon the grain size and composition of grains is extremely important in the classification and interpretation of paleosol units. The grain size distribution of a sediment or soil unit is one of the basic descriptive features in sedimentology or soil science. In these analyses the grain size was expressed in terms of the scale proposed by J.A. Udden and C.K. Wentworth (Tucker, 1981), which is shown in Table E1 (Appendix E1). This scheme is based upon measurement of particle grain size in geometric SI units. However, the scale can also be converted into an arithmetic one by use of the logarithmic formula:

$$\phi = -\log_2 S \qquad (5.3)$$

where, phi (ϕ) is the logarithmic transformation of the grain size (S) measured in millimetres (Tucker, 1981) and the resulting scale is shown in Table E1 (Appendix E1).

The soil profiles (Fig. 5.7) show a rough appraisal of mean grain size for each soil horizon estimated in the field using a grain size comparison chart. In order to establish the true grainsize, four accurately weighed samples of clay taken from the original clay lens by R.F. Vaughan (pers. comm., 1990) were passed through a stack of standardised metal sieves calibrated in both SI and phi units (Appendix E1). Sieving is a common method of assessing the grainsize of unconsolidated deposits of sand grade and above (i.e. >63μm in particle diameter) in both sedimentology (Tucker, 1981) and soil science (Klute, 1986). The resultant weights of material within each grain size fraction were measured and converted into percentages of the total sample. This data is shown in Table E2 (Appendix E1) and has also been converted into a cumulative frequency distribution.

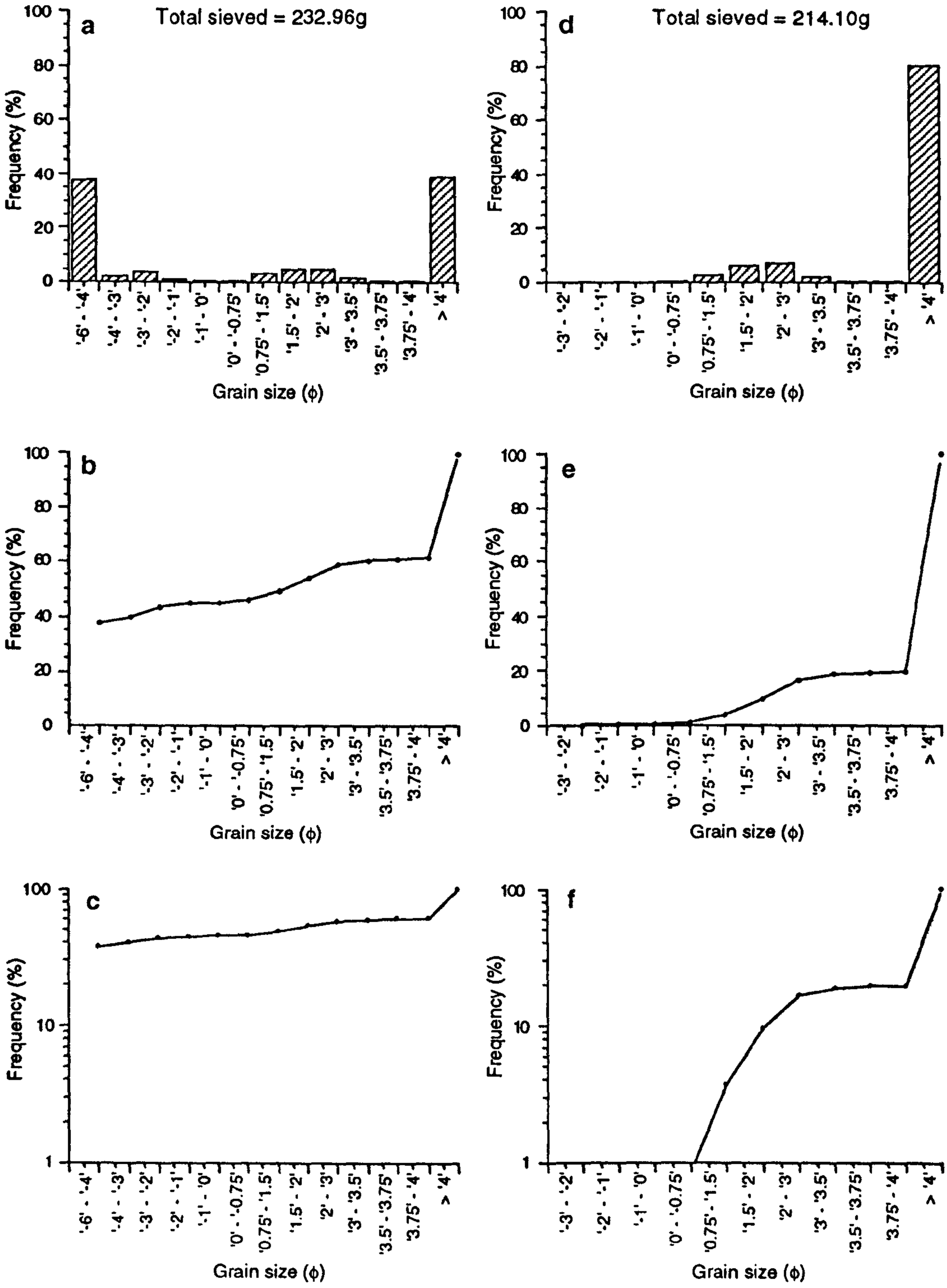


Figure 5.9. Histograms (a,d) and cumulate frequency charts (b,e : linear scale; c,f : log-scale), showing the results of a grainsize analysis carried out for the 'BC' (a-c) and '2Btg' (d-e) horizons of the Grey Clay paleosol.

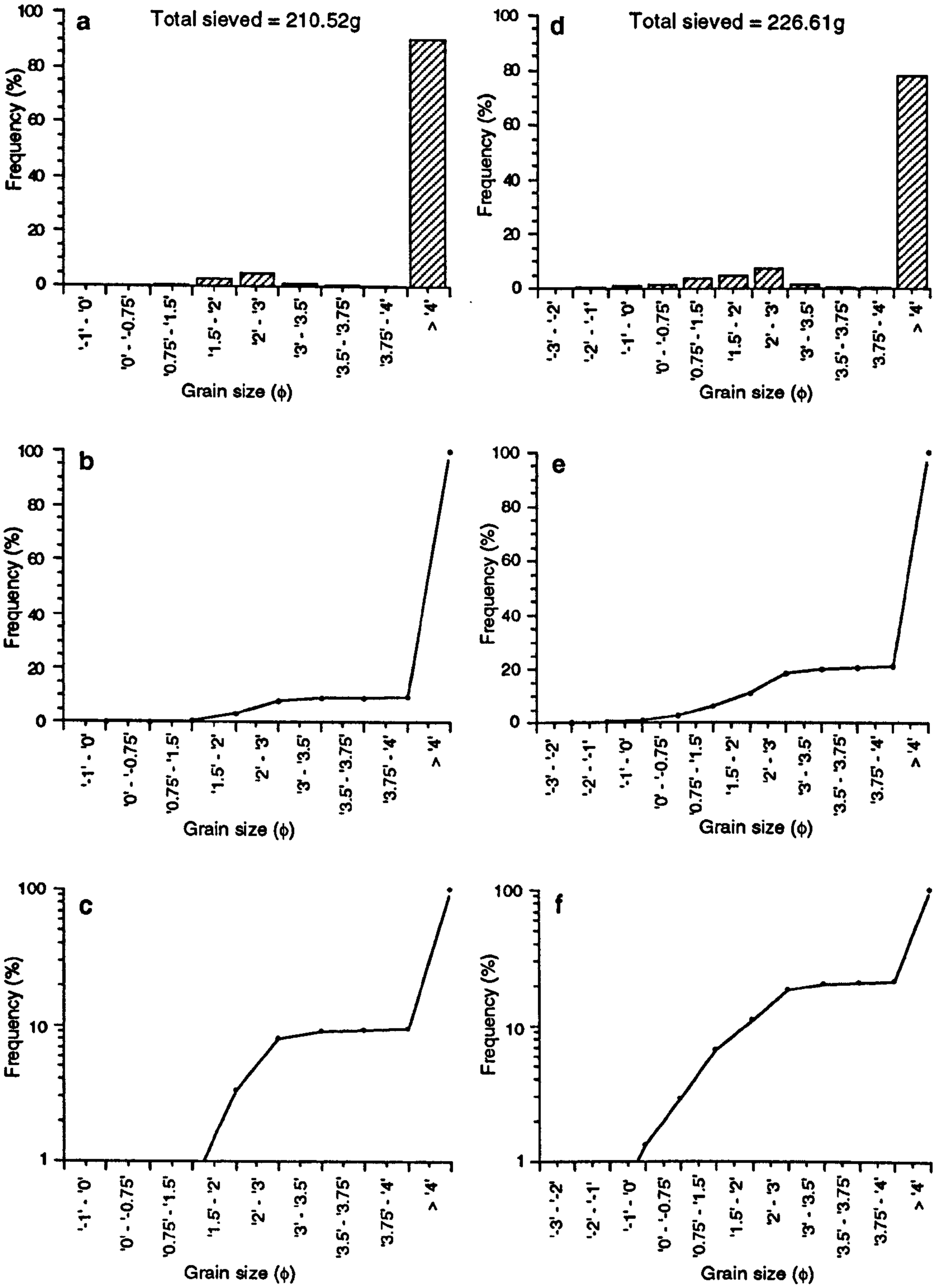


Figure 5.10. Histograms (a,d) and cumulate frequency charts (b,e : linear scale; c,f : log-scale), showing the results of a grainsize analysis carried out for the 'Btg' (a-c) and 'Aac' (d-e) horizons of the Green Clay paleosol.

The data has been represented graphically by plotting the frequencies (in terms of weight percentage or cumulative percentage) against the grain size fractions (in phi units) with grain size decreasing along the x-axis away from the origin. The resulting graphs for each sample are a histogram and smoothed frequency curve based on the first set of data and a cumulative frequency curve, and are shown in Figs. 5.9 (for Grey Clay horizons 'BC' and '2Btg') and 5.10 (for Green Clay horizons 'Btg' and 'Aac'). The cumulative frequency curve is plotted against a log probability y-axis.

Apart from the 'BC' horizon, all the graphs illustrate an extremely high proportion of fines, i.e. silt and clay (Figs. 5.9d-f and 5.10). None of the cumulative frequency curves approach a straight line plot, which would indicate that the distributions are normal or Gaussian (Tucker, 1981) and the smoothed frequency curves are not symmetrical but show a tail of fine material. This distribution is called a negative or fine skew (Tucker, 1981) and is reflected by the statistical tests carried out upon the data sets (see below). The 'BC' data shows a bimodal distribution, resulting from several large pebbles and one cobble of oolite found amongst the essentially fine-grained matrix. The poor quality of the resulting cumulative frequency curves meant that the usual method of ascertaining statistical parameters from the curve was impossible (for example, geometric mean, median and modal grain size, geometric standard deviation or the 'degree of sorting' of the data, and skewness or symmetry of the distribution; cf. Tucker, 1981). Instead the parameters were calculated by the purely mathematical 'method of moments' (Tucker, 1981), were each 'moment' corresponds to a meaningful parameter for grain size analysis. The formulae for the method of moments are:

$$\text{Mean (1st moment)} = \bar{x} = \frac{\sum fm\phi}{100} \qquad (5.4)$$

$$\text{Standard deviation (2nd moment)} = \sigma = \frac{\sum f(m\phi - \bar{x})^2}{100} \qquad (5.5)$$

$$\text{Moment coefficient of skewness} = \alpha_3 = \frac{\sum f(m\phi - \bar{x})^3}{100\sigma^3} \qquad (5.6)$$

$$\text{Mean-cubed deviation} = \alpha_3\sigma^3 = \frac{\sum f(m\phi - \bar{x})^3}{100} \qquad (5.7)$$

where f is the percentage fraction in each class interval of the total weight of sediment in each sample (Tucker, 1981). The arithmetic parameters are shown in Table 5.4. As is suggested by the plots for the Green Clay samples, the distribution is strongly negatively skewed toward the finer material (<63μm), with a skewness value (α^3) of greater than -0.30 (Folk & Ward, 1957). This is also the case with the 'Btg' horizon of the Grey

Bed description	Arithmetric mean (x)	Standard deviation (σ)	Coefficient of skewness (α_3)	Mean-cubed deviation ($\alpha_3\sigma^3$)
Bed B	0.56ϕ	4.70	-0.04	-4.33
Bed C	5.24ϕ	1.61	-1.84	-7.69
Bed D	5.65ϕ	1.09	-2.96	-3.83
Bed E	5.12ϕ	1.77	-1.75	-9.69

Table 5.4. Statistical parameters for the four sieved fractions removed from the Hornsleasow paleosol profile.

Clay. The bimodal distribution seen in Fig. 5.9a-c for the 'BC' horizon of the Grey Clay, has made the coefficient of skewness almost meaningless, with a value which suggests that the distribution is near symmetrical (Folk & Ward, 1957). This is also reflected in the value for standard deviation or sorting (σ) of the data, which at $\sigma = 4.70$ indicates that the distribution is very poorly sorted (Folk & Ward, 1957). The three other horizons show a lower standard deviation, which indicates that they are poorly sorted (Folk & Ward, 1957).

The arithmetic mean (x) of the strongly negatively skewed distributions is between 5-6ϕ, which corresponds to a grain size diameter of between 15-30μm or medium to coarse silt fraction (Table E1, Appendix E1). The value for the horizon 'BC' of 0.56f indicates a grain size diameter of around 0.75mm or coarse sand, but this is not a meaningful statistic.

In order to find the true mean grain size of the sub-sand fraction or 'fines', thirty-four 2g samples of clay were taken from the large blocks removed from excavation Site B. These were sampled from the same position as petrographic and petrological samples (see Fig. C2, Appendix C) and were labelled accordingly. These samples were analysed using the Elzone machine, a sedigraph which counts grain size distribution based upon the electrostatic conductivity of a solution made up of the individual sediment samples and the method of use is described in Appendix E2. The Elzone machine produced a printout giving the geometric mean, median, and modal grain size, and the geometric standard deviation. It also provides the arithmetic mean grain size and the arithmetic standard deviation upon that value. These measurements are expressed in SI units and as no statistical analyses were carried out upon the fine fraction alone, they were not converted into phi units. The average of three separate analyses are shown in Table E4 (Appendix E2).

The analysis reveals that the grain size of the fine fraction does not vary significantly over the clay lens and is essentially within the very fine silt and clay fractions (i.e. between 4-6μm or 7-8ϕ; Table E1, Appendix E1). However, it does appear that the grain size does increase slightly from 2.6-3.8μm within the Green Clay to 4.2-6.0μm within the lower Grey Clay units and this might reflect the different clay mineralogy's (see below). The grain size is also lower within the carbonate-rich 'C' horizons and may therefore be measuring individual grains of micrite. The standard deviation is quite high at between 1-2μm, and suggests a low degree of sorting, but these two parameters cannot be directly compared, as the Elzone measurement is of standard deviation upon the mean only and is not expressed in phi units (Folk & Ward, 1957).

Overall the grainsize analysis indicates that the paleosol units are essentially clay-rich, but are also quite poorly sorted. This poor sorting could reflect two main palaeoenvironmental conditions:

(1) The soils are weakly developed and have formed upon an essentially fine-skewed sediment, such as an abandoned river channel or crevasse splay deposit.

(2) The soils are very well developed and have been mixed by the actions of bioturbation and pedogenesis within the soil.

Clearly, it would be difficult and dangerous to attempt to resolve which, or whether both, hypotheses are correct from grain-size analyses alone.

5.3.4. Calcium carbonate content

A very rough approximation of the amount of calcium carbonate within a soil can be determined in the field by the standard application of a drop of 10% hydrochloric acid. The varying degrees of reaction of the soil to the acid, in the form of liberation of CO_2, have been categorised in a standard table given in Retallack (1990, table 3.2) and vary from no reaction at all, indicating a noncalcareous soil, to a vigorous reaction entailing frothing and doming of the acid drop, which clearly indicates a strongly calcareous soil. The test was carried out upon each soil horizon in turn and the results are shown in Table 5.3. The scale is quite subjective and, as the calcium carbonate within the soil horizons at Hornsleasow and particularly in the upper Green Clay, often occurs as discrete areas, such as carbonate nodules and relict clasts of oolite (there is no continuous 'calcic' horizon though). This approximate study was also supplemented by petrographic and chemical determination of calcium carbonate abundance.

5.4. Root traces

The presence of root traces within a bed, are the best indication that the horizon was once a soil. The top of the soil horizon can be defined as the surface from which the roots emanate (Fig. 5.5a) and are most densely concentrated (Retallack, 1990). The soil profiles (Fig. 5.7) and field sections (Figs. 5.1 & 5.8) show many iron-stained sinuous features which pervade from the top surface of the clays and these are identified as rootlet traces (Metcalf *et al.*, 1992). However, it is important to be able to distinguish root traces from other trace fossils commonly seen in sediments, such as burrows.

5.4.1. General root characteristics

The most obvious distinctions of fossil rootlet traces are the fact that they tend to taper and branch downwards, and are almost always orientated vertically (Retallack, 1990). They may also show numerous extremely fine lateral projections or root-hair traces, and sometimes exhibit internal anatomical detail (FitzPatrick, 1993; see section 8.4.2). The rootlets within the Hornsleasow soil profiles are most concentrated within the 'A' horizon of the Green Clay (Fig. 5.7), were they constitute about 3-4 rootlets per square decimetre. They are also seen to pervade the whole of the Green Clay layer, but

generally do not extend into the underlying Grey Clay (Fig. 5.7) and never penetrate the karstic limestones. Their confinement to the upper horizons suggest that the Green Clay was an aerobic soil, and as roots rarely penetrate permanently waterlogged parts of soils (Retallack, 1990), that this layer was fairly well drained. That they do not pass into the underlying Grey Clay suggests that this lower horizon remained waterlogged after silting-up of the pool and that the top of the Grey Clay ('O' horizon on Fig. 5.7) might corresponded to the palaeo-watertable depth in Green Clay times.

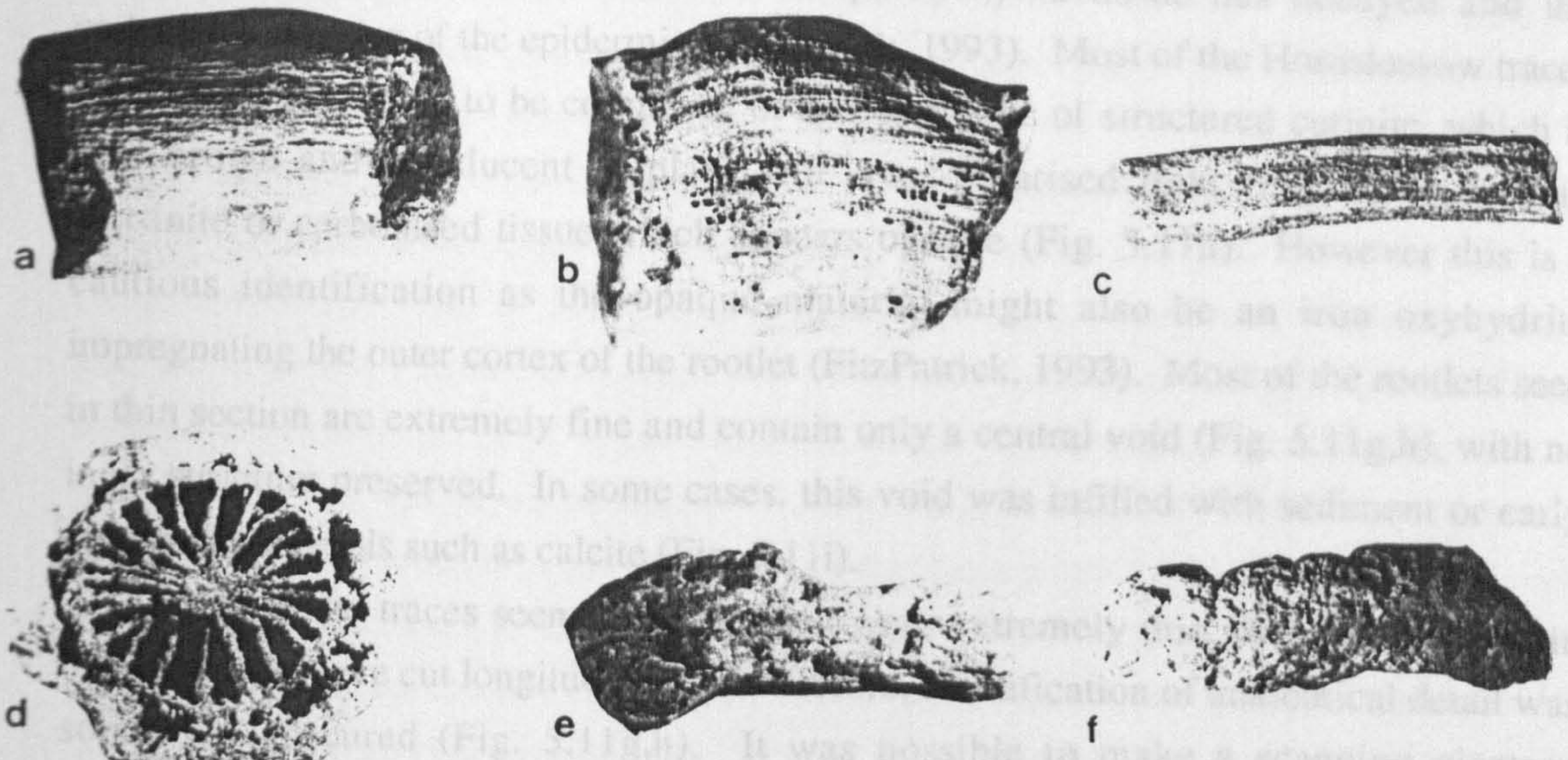
The rootlets are generally tubular and irregular in cross-section, with a width of between 2-5 millimetres. They are generally quite short, reaching lengths of between 10-45 millimetres, and tend to taper and branch downwards in the soil. Root types are fairly cosmopolitan and tend to be environmental specific, rather than indicating taxonomic affinities (Retallack, 1990). However, the general shape and form of the Hornsleasow rootlets are similar to the *Equisetites* root beds of the Yorkshire Scalby Formation (Hemingway, 1974) and they can be considered to be lycopodacaeous in origin (Metcalf *et al.*, 1992). The surface of the Green Clay was apparently covered in thick vegetation, although the overlying transgressive sands at Hornsleasow have all but removed the evidence except within the low-lying hollows upon the karstic surface.

The root traces in the Green Clay are mostly preserved as limonite ($FeO[OH].nH_2O$) stains and in the form of ferruginous rhizocretions (section 5.4.3). In slightly reducing conditions, iron oxide will often impregnate the surface of old root passages and following decomposition, iron 'pipes' will form (FitzPatrick, 1993; section 5.4.3). These are fairly stable and can remain unaffected for long periods of time or may be occupied by later generations of rootlets. Many of the larger root traces within the Green Clay are essentially composed of ferric oxide coatings or pipes where the original organic rootlet has rotted away (Fig. 5.11a-d). In most instances these pipes are simply hollow. However, some contain carbonised organic material or a sediment infill, whilst others show preferential impregnation of the original internal structure and these appear as radiating spurs or 'septa' (Fig. 5.11d). In extreme circumstances the rootlet has been completely replaced by limonite and the coating has become a ferruginous concretion (section 5.4.3; Fig. 5.11e,f).

5.4.2. Root structure in thin section and under the scanning electron microscope

The internal anatomy and preservation of the Hornsleasow rootlets was studied during the examination of the petrographic thin sections of the palaeosol horizons (section 5.5). Rootlets are anatomically simple and consist of a central decay resistant core made up of tracheid cells, called the 'stele', surrounded by a wide mass of bubbly tissue, known as the 'parenchyma cell tissue', which decay rapidly and the outer tough skin or 'epidermis' (Retallack, 1990). Typically in thin section root traces consist of a central carbonised

Figure 5.11. Root traces in the Hornsleasow paleosols. Iron pipes and ferruginous rhizocretions (a-f): (a) G. 89601, striated and hollow iron pipe, 9.3mm long; (b) G. 89603, sediment-infilled iron-pipe showing original root structure, 7.2mm long; (c) G. 51350, fine limonite replaced root-trace, 8.3mm long; (d) G. 89604, interior view of iron-pipe showing limonite-replaced root septa, 5.4mm in diameter; (e) G. 89700, ferruginous rhizocretion, 16.3mm long; (f) G. 89701, ferruginous rhizocretion, 19.5mm long. Organic root traces in petrographic thin section (g-i): (g) hollow, cutinite-sheathed rootlet in section HQS9-2(1), plane-polarised light (field of view 1.2mm); (h) cutinite (brown) and inertinite (opaque) root-trace in section HQS9-2(1), plane-polarised light (field of view 2.6mm); (i) calcite-infilled hollow organic root-trace in section HQS9-1, cross-polarised light (field of view 3.2mm). Pyritised rootlet (j) 3.1mm in length, inset (k) shows preserved cell-structure (field of view 320μm).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'streak', surrounded by a void where the parenchyma tissue has decayed and the carbonised remains of the epidermis (FitzPatrick, 1993). Most of the Hornsleasow traces appear in thin section to be composed of an outer zone of structured cutinite, which is pale brown and translucent in plane and cross-polarised light (Fig. 5.11g,h), and inertinite or carbonised tissue which appears opaque (Fig. 5.11h). However this is a cautious identification as the opaque material might also be an iron oxyhydrite impregnating the outer cortex of the rootlet (FitzPatrick, 1993). Most of the rootlets seen in thin section are extremely fine and contain only a central void (Fig. 5.11g,h), with no inner structure preserved. In some cases, this void was infilled with sediment or early diagenetic minerals such as calcite (Fig. 5.11i).

All of the traces seen in thin section were extremely thin (less than 100μm in diameter) and were cut longitudinally. Therefore, identification of anatomical detail was somewhat obscured (Fig. 5.11g,h). It was possible to make a scanning electron microscope study of the anatomy of a much larger pyritised rootlet recovered from the sieved fractions (Fig. 5.11j). The early pyritisation of the rootlet has prevented tissue compaction and this tiny fragment shows beautifully preserved three-dimensional cellular structure within tangentially broken sections (Fig. 5.11k). The study revealed an anatomically simple internal structure, which although not taxonomically diagnostic, appears to be largely composed of the decay-prone parenchyma cells, suggesting that pyritisation occurred early on in the decay process (section 5.9.4).

5.4.3. Ferruginous rhizocretions

Many of the iron pipes discovered during the sieving of the upper clays at Hornsleasow can be considered to be ferruginous rhizocretions (Brewer, 1976). Some of the pipes features show a surface striation (Fig. 5.11a-c), which is probably a cast of the original organic structure of the root epidermis. Others, however, exhibit a nodular and heavily encrusted appearance (Fig. 5.11e,f), which is almost certainly related to concretion formation. Most of the rhizocretions reach a width of 10-15 millimetres, although the original rootlet may have been much finer. It is likely that many of these roots would have become so encrusted with iron oxides that the original root would have died and decayed. Some of the rhizocretions contain carbonised organic material within their core.

The rhizocretions are composed of nodular ferric iron oxides, in particular haematite (Fe_2O_3) and limonite. Ferruginous rhizocretions form within the soil (Brewer, 1976) and indicate that there was free passage of well-oxygenated water (Fig. 5.12; Retallack, 1990). Rhizocretions form because of special localised conditions which surround and are created by the roots. The limonitic yellow mottling, limonitic nodules, rhizocretions and root traces abundant within the Green Clay 'A' horizon at Hornsleasow (Fig. 5.7) are most likely to be caused by the difference between the conditions prevalent

in the surrounding soil matrix and those within the micro-environments surrounding the rootlets and are related to the mobilisation and subsequent oxidation of reduced iron (Fe^{2+}) within the acidic rhizosphere (Fig. 5.12).

The uptake of water by rootlets, is often accompanied by an uptake of nutrient cations, such as iron, calcium, magnesium, potassium and sodium from the soil (Retallack, 1990). This is facilitated by the roots exuding substances which can release and mobilise the nutrient ions from their original minerals (Richards, 1974). These include organic acids, such as caffeic acid, which can liberate Fe^{2+} from the ferroan calcite lattice by hydrolysis (equation 4.1, Fig. 5.12; Trudgill, 1985), this reaction also produces CO_2. Ferrous iron can also be fixed and incorporated into the structure of some organic molecules, such as the organic acid, ethylene diamine tetracetic acid (EDTA), by a process known as chelation (Trudgill, 1985). Also, as the roots respire they give off CO_2, which in turn is converted into carbonic acid (equations 4.2 & 4.3), increasing the aggressiveness of the rhizosphere toward minerals (Fig. 5.12; Richards, 1974). The net effect of the root action is to deplete the surrounding sediment of the nutrient ions, and this depletion is most prevalent in the growing tip of the rootlet (Russell, 1977).

However, the rhizosphere is not usually uniformly acidic or reducing, and most of the time conditions are at the near neutral redox potential which supports the normal aerobic respiration and growth of the roots and bacteria which inhabit the zone (Richards, 1987). It is only when conditions are unstable or fluctuate over short periods of time that the uptake of nutrients and precipitation of other minerals can induce rhizocretion formation (Brewer, 1976). It is thought that repeated cycles of wetting and drying of the rhizosphere are responsible for concretion formation and growth. For instance, heavy rainfall will produce temporary waterlogging of the soil and will induce the reducing and acidic conditions which favour ferrous iron mobilisation (Fig. 5.12). It may also increase rootlet productivity, especially after long periods of drought (Retallack, 1990). Excess iron that has been mobilised in the drab ferrous state within a reducing wet rhizosphere can then be oxidised near to the rootlets to yellow and red ferric oxides to form concretions during drier periods (Fig. 5.12; Brewer, 1976).

The formation of the limonitised root traces and associated oxidised yellow limonitic mottling in the Green Clay units may also be purely diagenetic after pyrite (section 5.9.4). However, the presence of the ferruginous rhizocretions within the Green Clay would suggest that most of the oxidation occurred during the life of the soil.

5.5. Soil structure

In the field the Hornsleasow soil units appear to be largely featureless beds of massive clay, especially in comparison to the striking cross-bedding exhibited in the Chipping Norton Formation limestones. Soils and paleosols are not structureless, however, as

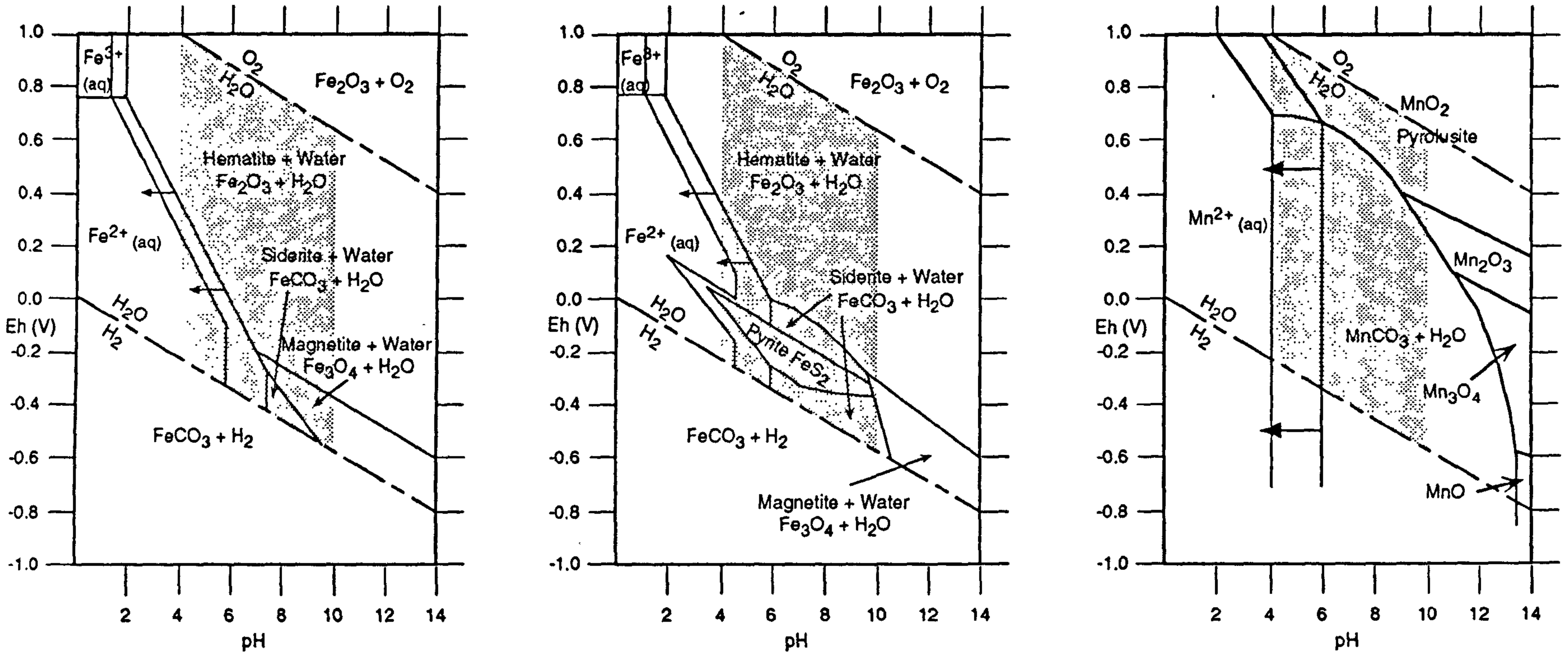


Figure 5.12. Stability fields in aqueous solutions at one atmosphere pressure and 25°C for (a) haematite, magnetite and siderite, (b) iron oxides, sulphides and carbonate, and (c) common sedimentary manganese minerals. Natural freshwater systems are shaded.

they exhibit a combination of features inherited from their parent material (such as relict bedding and relict or 'skeleton' grains) and structures which have developed during soil formation. There is an incredible amount of technical jargon incorporated into the soil science descriptions of such structures (cf. Brewer, 1976, FitzPatrick, 1993) and so I made an attempt to keep terminology to the barest minimum in this work.

Common structures seen in modern soils are the open network of pores and cracks, surrounding clumps of soil material or skeleton grains of rock (Fig. 5.5b). Burrows and root traces are also fairly common features. Other features include hardened lumps of soil and nodular material chemically and mineralogically distinctive from the original rock and the soil matrix. The soil matrix itself is usually composed of a fine grained mixture of clay and amorphous iron and aluminium oxides. Organic material is also abundant in some soils and some may exhibit strong colour variation or mottling.

The features listed above are extremely important in determining the structure of a fossil soil, even though some, such as the pore space and organic matter content, may be altered by burial, compaction and diagenesis. A full petrographic study of the four large soil blocks removed from site B was carried out and the various features of the Hornsleasow paleosol units are described in the following sections 5.6-5.9. In order to study some of the more finer structures, such as microfabric and identification of skeletal and pedogenic clasts, microtome thin sections were made of 24 samples of clay taken from regularly spaced intervals and specific soil horizons upon the four soil blocks (Fig. C2, Appendix C2). Unfortunately the sectioning proved to be an extremely difficult process (Appendix C2), and of the original 24 samples, only fifteen thin sections were useful. These were described petrographically under natural daylight, and in plane and cross-polarised transmitted light at x40 and x100 magnification under the petrographic microscope. Full descriptions of these thin sections with explanations to various features seen in petrographic section are given in Appendix C5.

5.6. Peds, cutans and pore space

Peds are the aggregated clumps of loose soil matrix which lie between the cracks, roots and pore space in a soil horizon (Retallack, 1990), they are commonly called 'clods of earth' by gardeners and farmers. Peds range from sand-sized particles to large clumps of soil, and may be composite in form (Fig. 5.5b). Peds usually have a distinctive shape and form and are classified according to their size, shape and angularity (Retallack, 1990, fig. 3.11). The most common forms of peds are 'platy', 'blocky' and 'granular' or 'crumb' peds (Retallack, 1990). Platy peds form in weakly developed soils and appear to follow relict bedding. Blocky peds are irregular and almost equant in form (Fig. 5.5b) and have angular or subangular interlocking faces, depending upon the amount of erosion of the surface of the ped which has taken place. Blocky peds are common within clay-

rich soils and form when the soil is subjected to periods of wetting and drying (Retallack, 1990). Granular or crumb peds are small, equidimensional peds which are often found in the upper portions of well-bioturbated and rooted 'A' horizons. They form from the constant churning of the soil matrix or from deposition of faecal pellets (Retallack, 1990). Another type of ped structure is the large-scale prismatic or columnar peds seen in extremely arid soils, which are elongated with flat or domed top surfaces. These form by repeated wetting and drying of swelling clays, such as smectite (Retallack, 1990).

The pore space which define the peds are probably the most obvious features of modern soils. The pore system is the part of a soil which contains the soil atmosphere and/or moisture. 'Pores' are defined as the discrete voids within the matrix and 'pore-space' as the continuous void phase throughout the matrix (FitzPatrick, 1993). Common types of pores include small irregular pockets or 'vughs', and approximately spherical pockets or 'vesicles' (Fig. 5.5b; Retallack, 1990). The origin of vughs is unknown, but that of vesicles appears to be related to soil CO_2 formation as tiny gas bubbles (FitzPatrick, 1993). Other types of pores include burrows, root passages and planar, sinuous pores which appear to be related to shrinkage and cracking.

In thin section the spatial distribution and organisation of the soil is known collectively as 'soil structure' (FitzPatrick, 1993) and is a way of describing the relationship between pores, pore-space and ped forms. The structure of a soil in thin section can be described as complete, incomplete or massive, which separates soil which has well defined ped forms surrounded by continuous pore-space from those with incomplete and no continuous pore-space at all (FitzPatrick, 1993). There are many types of soil structure and a classification scheme has been attempted by FitzPatrick (1984, 1993) which defines classes of structure based upon the dominant type of arrangement of ped and pore system seen in thin section.

The most commonly observed structures in the Hornsleasow thin sections are:

(1) *'Angular blocky'*. This is were the peds form angular clods, and there is are continuous pore-space between peds and many discrete pores within the peds. It often occurs in the B horizons of fine-grained soils and suggests shrinkage and cracking during seasonal wetting and drying (FitzPatrick, 1993).

(2) *'Coat and bridge'*. This is were fine material forms thick coatings and bridges between sand or silt sized grains, in this structure peds are not usually formed. The pore system between the grains may be made up of irregular and discrete pores. In the field this can be confused with the leached 'single grain' soil structure, but in thin section the latter is seen to have frequent and connecting pore-space and no thick coatings of fine material. In the Hornsleasow paleosols this is seen in the more arenaceous-rich soils and upon the top surface of the karstified limestones (section 4.4.2). Coat and

bridge usually forms in sandy horizons, where organic matter and fine material is washed between the grains (FitzPatrick, 1993).

(3) *'Crumb or granular'*. This is the equivalent to the crumb and granular peds described above, and is characterised by small irregular shaped peds which are totally separated by the intervening pore-space. Pores are rare within individual peds, which are loosely bound together by organic matter, rootlets or clay. This type of structure is very common in the 'A' horizon of soils and can indicate bioturbation and root activity in an aerated soil (FitzPatrick, 1993). Granular structure can also occur within the upper horizons of soils rich in expandable layered clays, which are seasonally subjected to wetting and drying, a process known as 'self-mulching' (FitzPatrick, 1993).

(4) *'Irregular blocky'*. This is were peds are broken up into irregular shapes, with many discrete sinuous and bifurcating pores and frequent pore-space. It is similar in appearance and origin to angular blocky structure and occurs within the middle horizons of finely textured soils (FitzPatrick, 1993).

(5) *'Laminar'*. This is the thin section equivalent to 'platy' ped structure and is characterised by elongated and strongly aligned peds, with frequent horizontal continuous pore-space, which run parallel to original bedding planes. Discrete pores are rare within the platy peds and the structure is typical of weakly formed soils (Retallack, 1990, FitzPatrick, 1993).

(6) *'Massive'*. This is were there is a continuous soil phase, and peds are rare or absent. Pore-space is limited and discrete pores are usually rare. This occurs within finely textured soil horizons and is particularly common in lower horizons were ped formation is inhibited by the weight of overburden and/or from being permanently waterlogged. Discrete pores form by gas release and occasional drying (FitzPatrick, 1993). This structure is also common in compacted paleosols.

(7) *'Composite'*. This is not a true soil structure class, but indicates a combination of two or more structural types. It is the commonest structure seen in the upper horizons of many soils and is caused by "two or more contrasting processes which are acting simultaneously or sequentially in response to seasonal changes" (FitzPatrick, 1993, p. 121-122).

The formation of pores within a soil is considered to be the initial stage of ped formation. For instance, the sinuous and planar pores which form upon shrinkage of the soil matrix may begin to bifurcate and the resulting cracks may spread outwards during prolonged desiccation. Eventually, the pore-space may become continuous and the pores will define an angular blocky ped structure. In many paleosols pore spaces can be destroyed by compaction and diagenetic petrifaction of the originally loose matrix, and the matrix may appear massive in thin section. However, in some paleosols former pore space can be indicated by recognising the modified surfaces of peds. The surface of a

ped or clod 'skin' often undergoes alteration at the contact between soil matrix and open pore-space, particularly if the pores are fluid-filled (Retallack, 1990). Within paleosols these would form irregularly planar features of alteration and can be detected in thin section. These relict ped skins are known as 'cutans' (Fig. 5.5b).

In detail, cutans are described by a complicated classification scheme, which provides the particular structure with a Latinate name indicating the composition and the suffix '-an' indicating that the structure is a cutan (Retallack, 1990). Cutans are usually irregular, thin and extremely fine-grained, whereas diagenetic veining which can be confused with original pore-space is usually thicker and coarser grained.

There are three categories of cutan formation, and these are illuviation, diffusion and stress cutans. Illuviation cutans form when material is washed into pore-space from above and are thus, composed of material different from the soil matrix (Retallack, 1990). They commonly show sharp contacts with adjacent soil peds and may be laminated in thin section, from successive depositional washings. Diffusion cutans are formed by progressive alteration inward from a pore-ped surface (Retallack, 1990). They have a sharp outer boundary and a diffuse inner alteration zone and are not composed of material strikingly different from the original ped matrix. Stress cutans are not easily distinguishable from surrounding soil matrix, unless studied in thin section. They are most easily seen in clay matrices, where ped surfaces have repeatedly sheared past one another during swelling and shrinking episodes. They are also called 'slickenslides', and in thin section are characterised by thin lines of highly anisotropic bright clay surrounded by a gradation into normal isotropic clay (Retallack, 1990; section 5.8). Slickenslides can also form in paleosols during compaction, as peds are crushed. Hence, by themselves slickenslides are not conclusive evidence that a clay paleosol underwent seasonal wetting and drying.

There are four types of cutans observed in the Hornsleasow paleosol horizons and these are:

(1) *'Argillans'*. These are clay skins to silt grade material. Most of the argillans identified in the Hornsleasow sections are slickenslides (section 5.8).

(2) *'Skeletans'*. These are veins of skeleton grains and are typical illuviation cutans. The ones that are seen in the Hornsleasow thin sections are in the main composed of quartz grains.

(3) *'Ferrans'*. These are more commonly called iron-stains and are composed of ferric oxyhydrites and iron oxides. Ferrans commonly form in fairly acidic, oxidising conditions and are often observed upon sand grains in well-drained sandy soils in humid climates. In the Hornsleasow soil they occur upon small granular clay peds and are observed as limonitic and haematitic coatings upon carbonate skeleton grains in thin section in both the soils and karstified limestones (Appendices C5-C6). They appear to

be alteration product weathering rinds and coatings of airborne dust and in this context, the Hornsleasow ferrans are considered to be diffusion cutans.

(4) *'Organans'.* These are clod-skins composed of organic matter, which has been deposited in cracks and pore-space and occur in many different soil horizons (FitzPatrick, 1993). Organans are another example of illuviation cutans and are fairly common in the surface horizons at Hornsleasow. Organic coatings upon grains show distinctive pattern of cracks which enables differentiation between these and other opaque coatings such as limonitic ferrans.

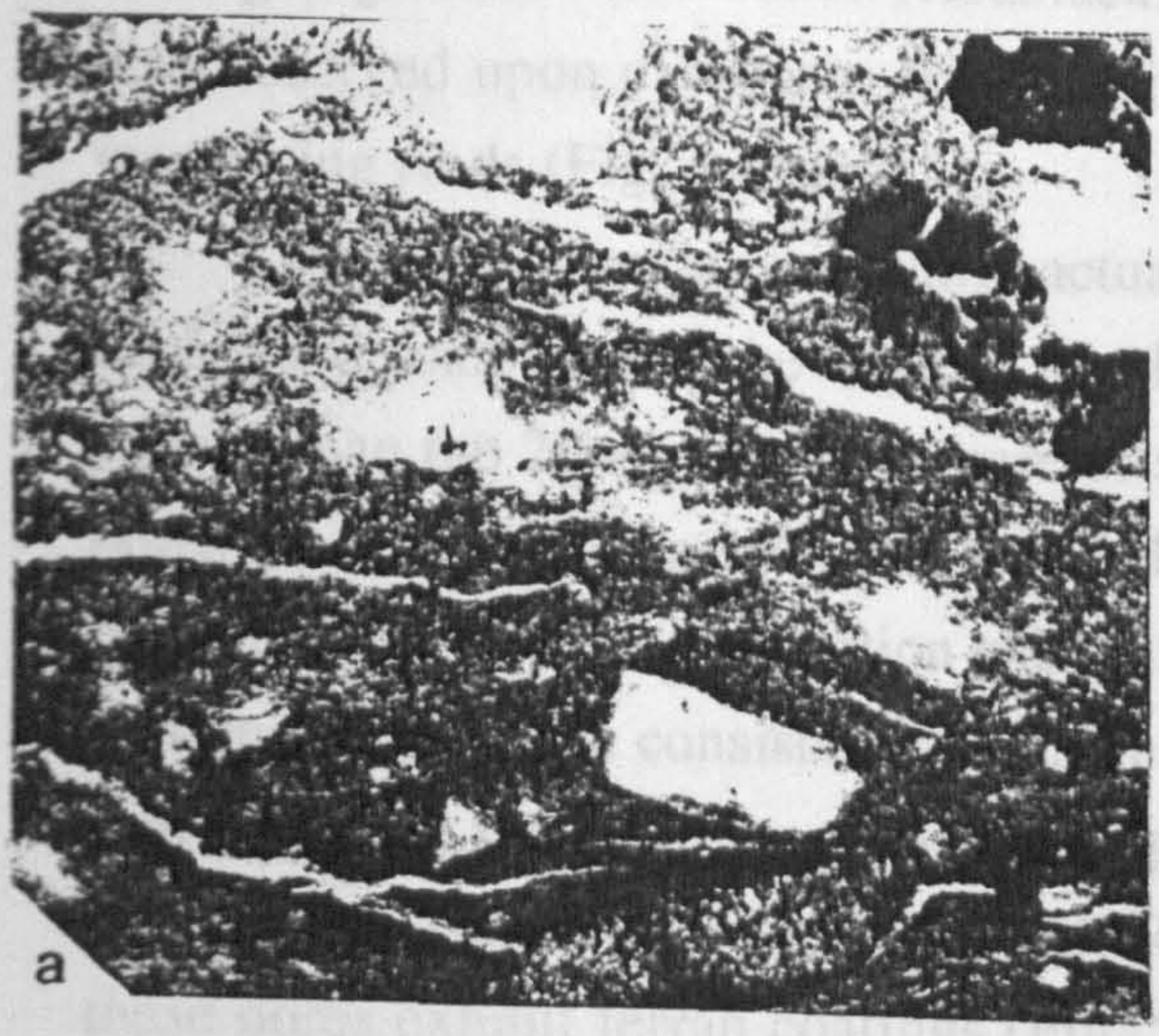
(5) *'Mangans'.* These are composed of manganese dioxide and appear as patchy fine coatings upon some of the larger oolite pebbles at the base of the Grey Clay ('BC' horizon). They commonly occur in soils which occur in moist or seepage conditions (FitzPatrick, 1993) and appear to indicate the waterlogging of the limestone hollow following blocking of the conduit during initial clay soil deposition (section 4.5).

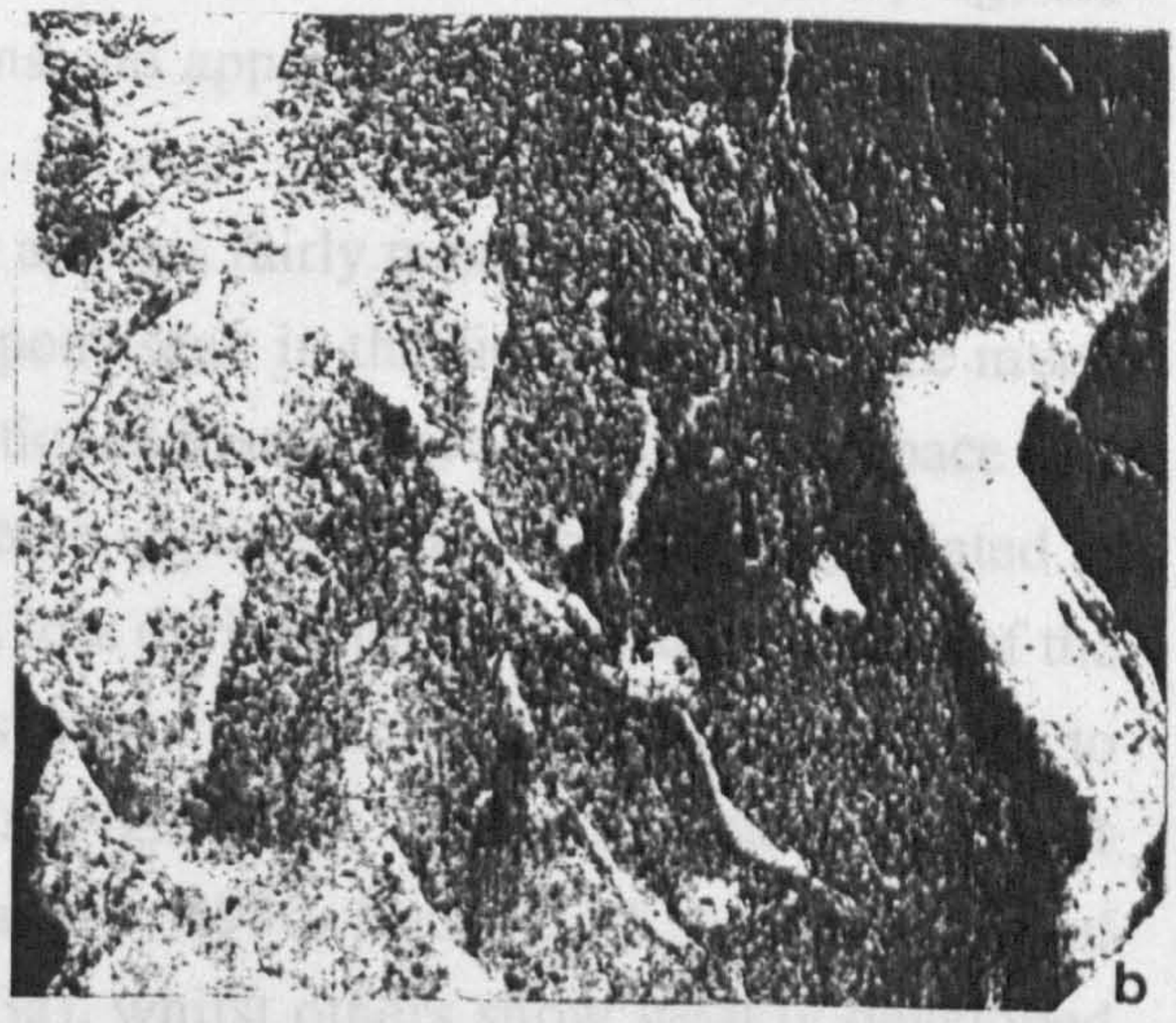
Full details of all these features seen in petrographic section and in the field are described in Appendix C5, for the Hornsleasow soil samples. Below is a summary of the structure of the individual soil horizons and some possible explanations as to their formation in the paleosols.

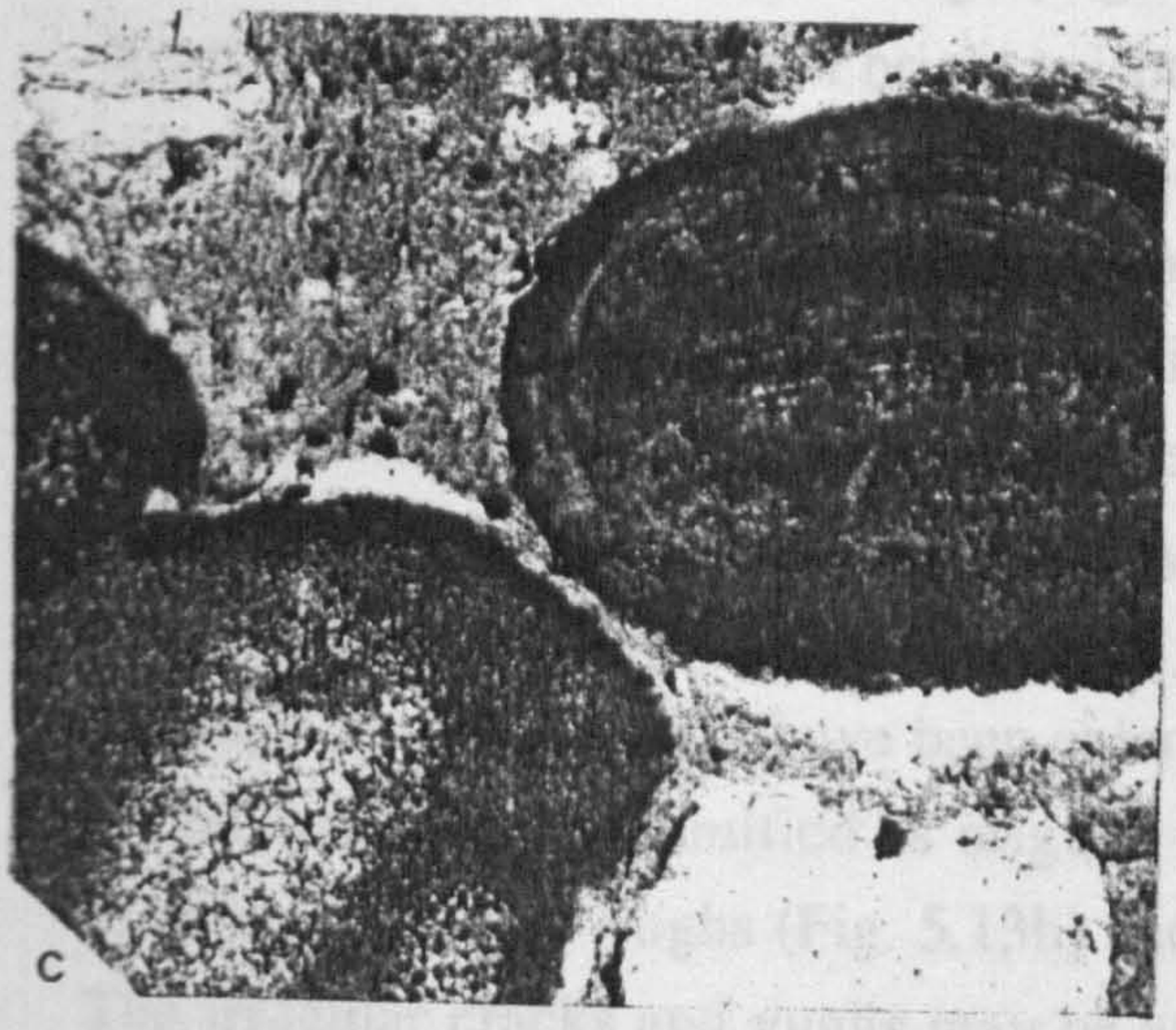
5.6.1. The Green Clay 'Aac' horizon

When initially dug, the Green Clay appears to be a fairly featureless and massive clay unit. In the upper horizon only the yellow mottling and the pervasion of root passages break up this monotonous unit. However, upon careful drying in the laboratory, the 'Aac' horizon becomes much more friable and is pervaded by fine cracks which break the surface up into medium to coarse (10-50mm) angular blocks (Fig. 5.8; cf. Retallack, 1990, figure 3.11, p.40). These features infer that the original soil horizon may have exhibited an irregular blocky ped structure. The sieving of the 'Aac' grain-size sample also reveals further ped development within the horizon, as up to 5% of the residue consisted of small (between 0.25-1.0mm) granular peds of ferruginized clay. The clay granules exhibit a yellow-stained surface and are strongly cohesive as peds compared to the rest of the clay fraction. These granular peds and associated ferran diffusion surfaces are probably associated with the yellow mottled regions of the surface horizon and confirm with the evidence from ferruginised root passages that this was a strongly oxidising region of the soil.

The ferrans are also observed in thin section (Fig. 5.13a,b) occurring as thin (1-2µm) zones of limonite and goethite granules surrounding open pore-space in the soil matrix. Organic matter may be incorporated within the ferruginous cutans which may impart the some of the dark colour (Fig. 5.13a; FitzPatrick, 1993). Granular peds cemented by organic matter and sesquioxide coatings are typical of surface 'A' horizons







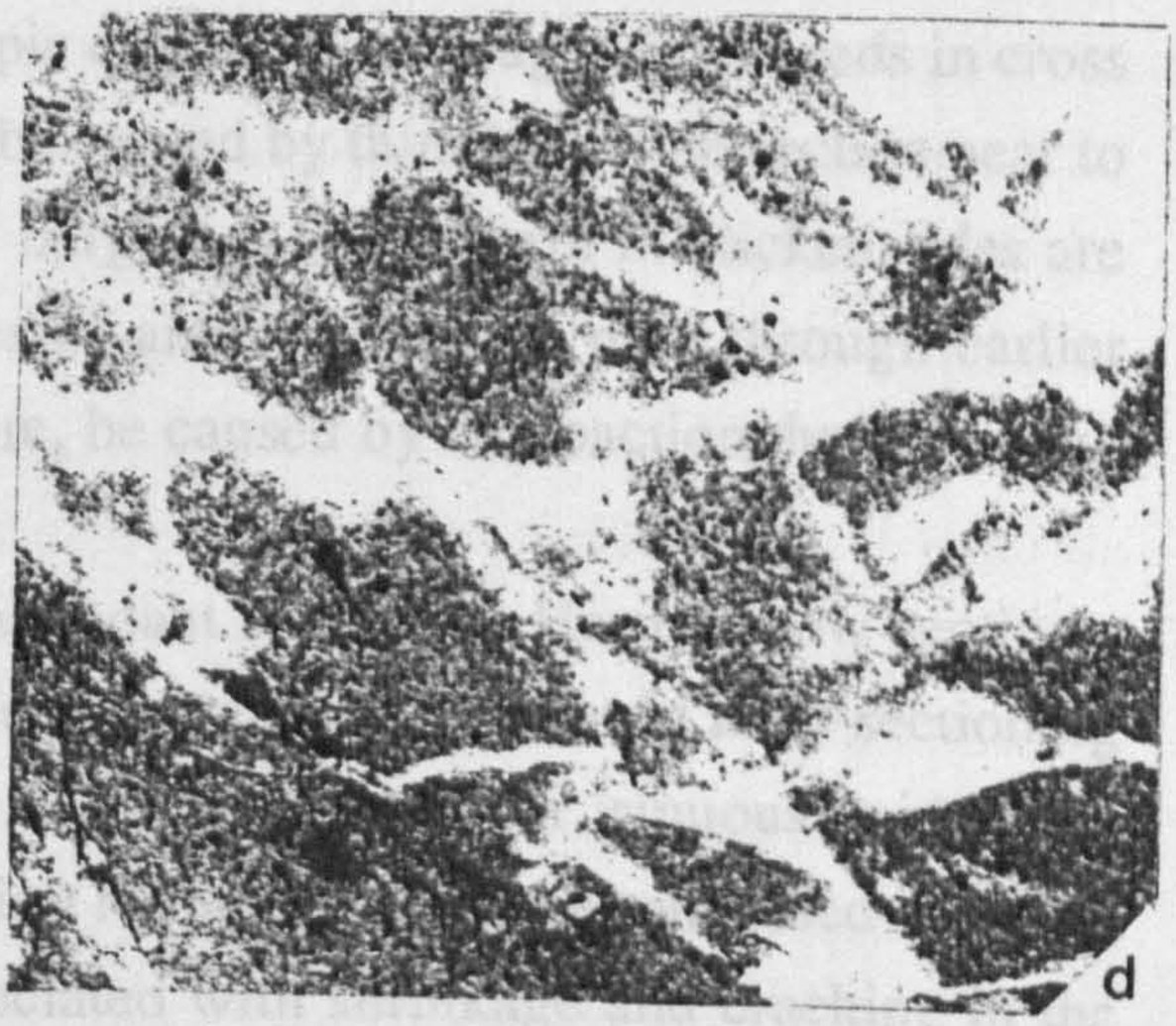


Figure 5.13. Photomicrographs of petrographic thin sections of the Green Clay paleosol: (a) section HQS10-1, plane polarised light; (b) HQS12-1, plane polarised light; (c) HQS9-1, plane polarised light; (d) HQS12-3, plane polarised light; (e) HQS11-2, plane polarised light; (f) HQS10-3, plane polarised light. Field of view 2.3mm, except in (c) 0.9mm. All sections were photographed under blue-filtered light.

undergoing active bioturbation (Retallack, 1990; Foos, 1991). Ferruginized coatings are also observed upon carbonate skeleton grains and appear to be part of strong micritic weathering rinds (Fig. 5.13a,c).

In thin section the soil matrix actually appears fairly massive in structure, but this could be a scale factor, the angular blocky peds seen in the field sections were much coarser than the 20mmx10mm square of sectioned material examined. Pore-space and pores are evident in the sections, although some apparent voids are probably related to poor and differential impregnation and matrix loss during preparation and grinding of the sections. Pore-space consists of thin, sinuous cracks which may bifurcate and join up to form very fine secondary angular blocky peds (Fig. 5.13a). Those in sections HQS10-1 and HQS12-1 are up to 4mm long (cf. Retallack, 1990, figure 3.11, p.40). Some of these pores exhibit ferran coatings (Fig. 5.13a), whilst others show weakly developed organans. The appearance of bright anisotropic clay along the edges of the peds in cross polarised light (Fig. 5.15c), may be actually be caused by thinning of the section near to cracks rather than the presence of argillans. Argillans in the form of slickenslides are seen in section HQS12-1, these bright lines of anisotropic clay pass through earlier matrix microfabric patterns and may therefore, be caused by compaction during burial, rather than be an original feature of the clay.

Discrete interpedal pores are fairly abundant within the Hornsleasow sections, although again some may have been enlarged or have resulted entirely from sectioning procedures. Those identified as original pores occur as irregular, sinuous voids (Fig. 5.13b), as irregular vughs (Fig. 5.13b) and tiny round or ovoid, smooth-sided vesicles. The irregular cracks and vughs may be associated with shrinkage and cracking in the clay-rich peds. The latter pores may be gas escape structures associated with the carbon dioxide surrounding the rhizosphere. Root passages are the only recognisable faunal features within the surface horizon in thin section, although other evidence suggests that the horizon was bioturbated. Although skeleton grains are fairly abundant (section 5.7.1), they do not form clastic dikes or skeletans in the sections and blocks studied. Veins of clastic material often indicate oligochaete (i.e. earthworm and related animals) activity and bioturbation (FitzPatrick, 1993).

5.6.2. The Green Clay 'Btg' horizon

This subsurface horizon is clay-rich and in the field displays a featureless massive structure. Very few root hair passages are seen and pore-space is limited. Upon laboratory drying, the clay horizon fragmented into coarse angular blocky 'peds' (Fig. 5.9). Sieving of the soil sample revealed less than 1% fine granular or crumb-like clay peds, these were not visibly iron-stained and may be pelletal in origin.

Thin sectioning of the Btg horizon was not always completely successful (Appendix C2) and many of the sections are fragmented during polishing. Large irregular interpedal voids with diffuse boundaries such as those seen in sections HQ S11-2 and HQ S12-2 were probably caused by poor impregnation and matrix or grain loss during preparation (Fig. 5.13d). Some of the sections are so poor that there is only a minute trace of soil matrix left on the glass slide (e.g. HQ S12-2, HQ S9-3 and HQ S10-3). In these cases much of the structure is inferred from the more successfully sectioned slides (Appendix C5).

The structure seen in the thin sections ranges from fairly massive to incomplete or composite irregular and angular blocky ped development. Much of the clay-rich areas have particularly low porosity (Fig. 5.13e), whilst areas rich in skeleton and sand grains possess more continuous pore space (Fig. 5.13f). Most of the sections show blocky ped development (Fig. 5.13d), although some of the fine crack-like pore space are now closed giving the impression of a completely massive structure, for example in HQS11-2 (Fig. 5.13e). The blocky ped structure can be seen very clearly in some of the better preserved sections when held up to the light and appears to be upon a very fine scale with peds less than 5mm, although some are up to 10-20mm in size (Retallack, 1990, figure 3.11, p.40). Most of the cracking in the clay-rich matrix shows evidence of bifurcation and triple-point nucleation of cracks to form the angular blocky peds (Fig. 5.13d), this kind of pore space development is typical of shrinkage within periodically wet middle and lower horizons of fine-grained soils (FitzPatrick, 1993). Interpedal pores are generally similar to those described in section 5.6.1., although many small crack like discrete pores are probably incomplete shrinkage cracks.

Again no burrows were recognised in the thin sections or within the sample blocks and the horizon does not appear to have been strongly bioturbated. Rare root passages do pervade the layer and have been identified in the thin sections (Fig. 5.13f). Disseminated organic matter is quite common in the sections and illuvial organans have been recognised in some sections. Other types of cutans are much rarer in the 'Btg' horizon of the Green Clay. Some of the peds may have diffuse regions of iron mottling at their surfaces and opaque-rich weathering rinds occur upon carbonate skeleton grains (Fig. 5.13f), but diffusion ferrans are not well developed on most ped surfaces. Clearly defined anisotropic lines of bright clay occur in section HQS11-2 (Fig. 5.15f) and do appear to be associated with angular blocky ped surfaces. This would again suggest some degree of shrinkage during seasonal precipitation variations in this subsurface horizon.

5.6.3. The Grey Clay 'Oe' and 'E/B' horizons

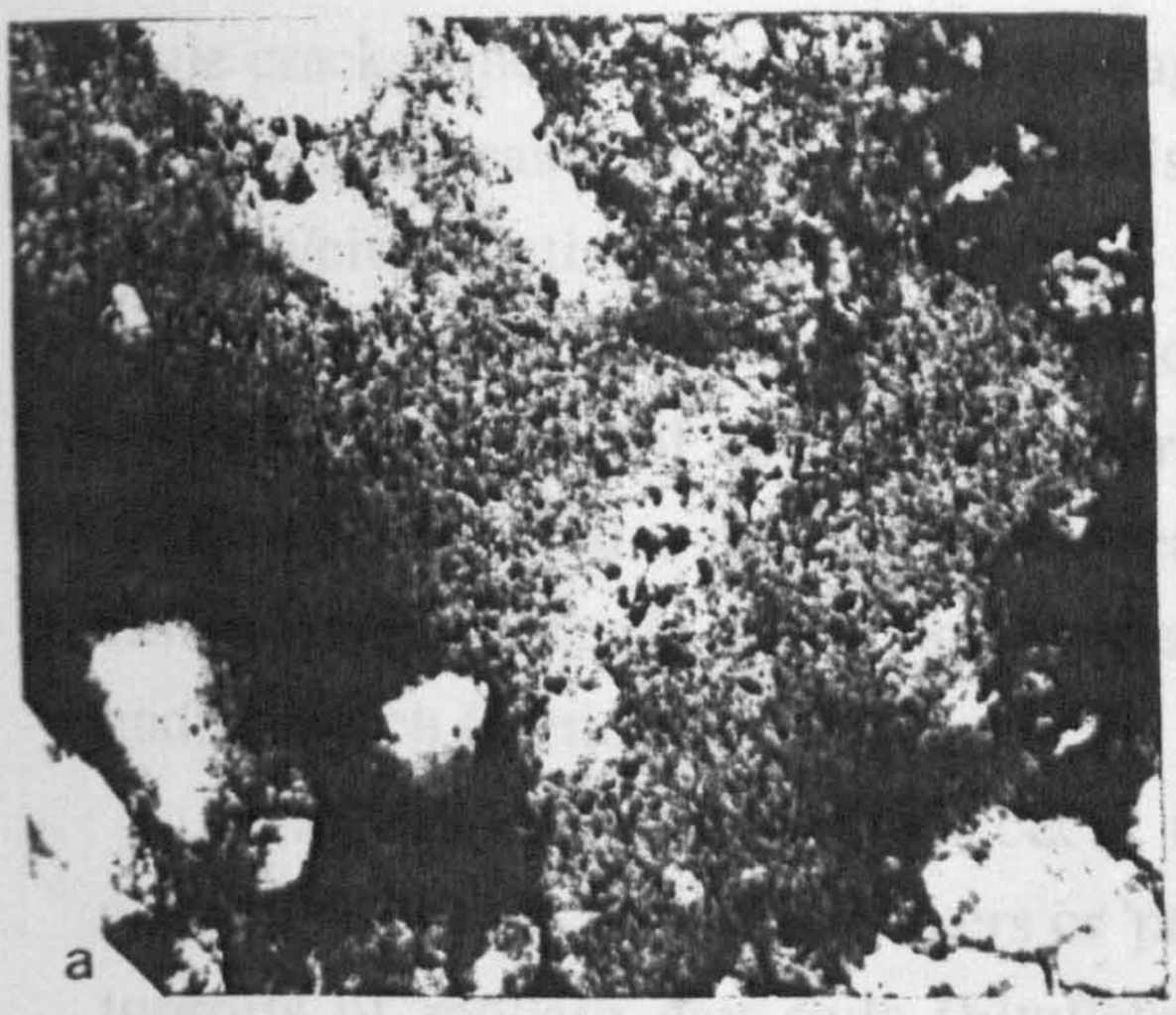
The thin organic rich layers at the interface between the Green and Grey Clay soil units, is largely massive and unstructured in hand specimen. The large carbonised pieces of organic matter tend to fragment into a fine black dust upon desiccation and the intervening clay is mainly massive. Root structures are not seen to pervade from the organic horizon and no burrows were noticed in the field.

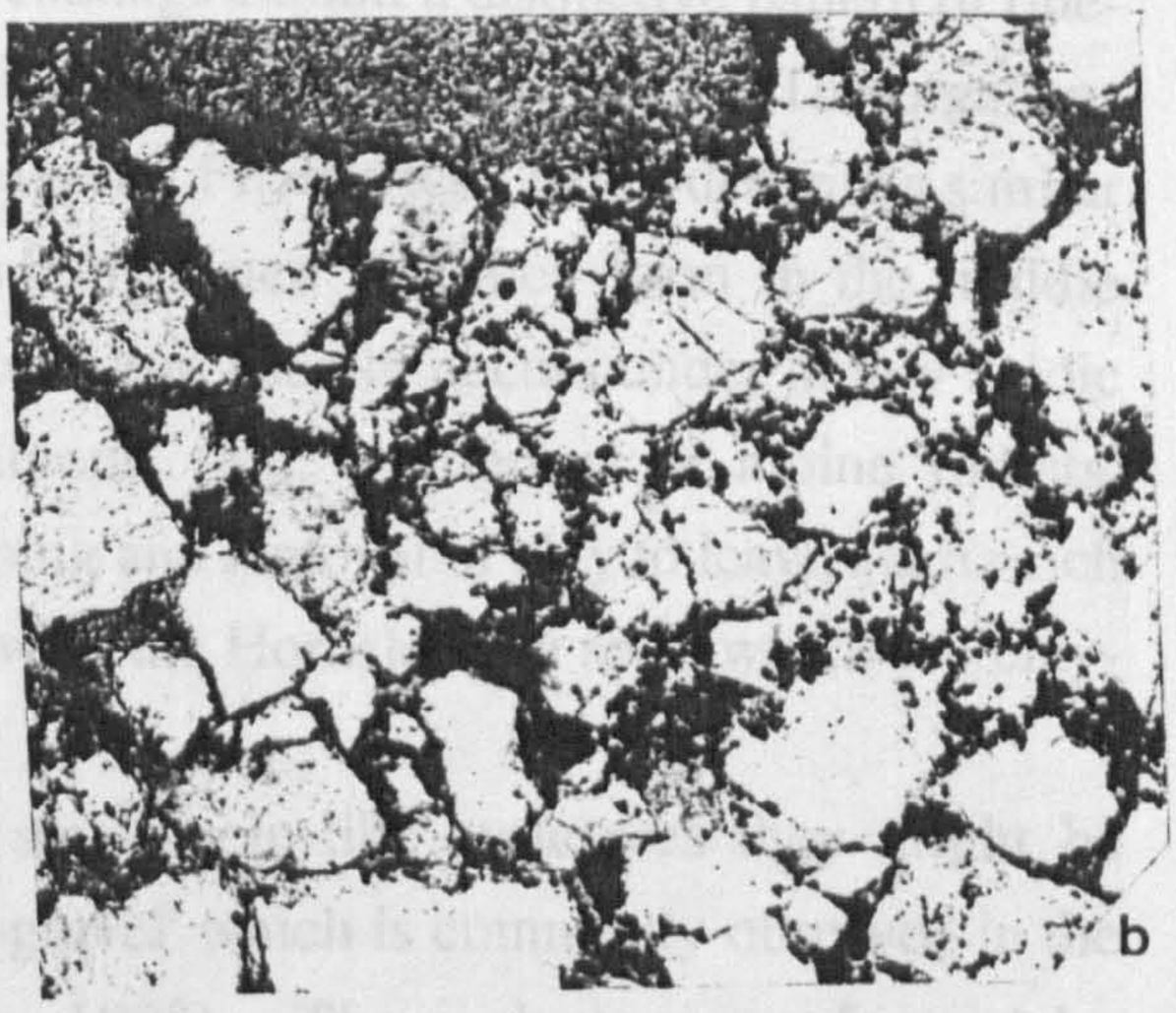
Preparation of the thin sections was again fairly difficult, and only HQS12-4 was sufficiently impregnated to observe medium and fine scale structures in the soil. Held up to the light, the thin section shows a system of discrete sinuous crack-like pores which probably originated from incomplete early stage desiccation and shrinkage in the soil (FitzPatrick, 1993). Up to 50% of the soil matrix has been lost in the section and grinding has enlarged some of the original pores space. Under the microscope, both the poorly preserved HQS10-4, and section HQS12-4 (Fig. 5.14a), exhibit a massive or coherent structure in clay-rich areas. Continuous pore space is rare or absent, although occasional discrete three-limb shrinkage cracks are seen in HQS12-4 (Fig. 5.16a), indicating incomplete blocky ped formation. Discrete irregular pores are also present, but there is mainly a continuous interpedal soil phase.

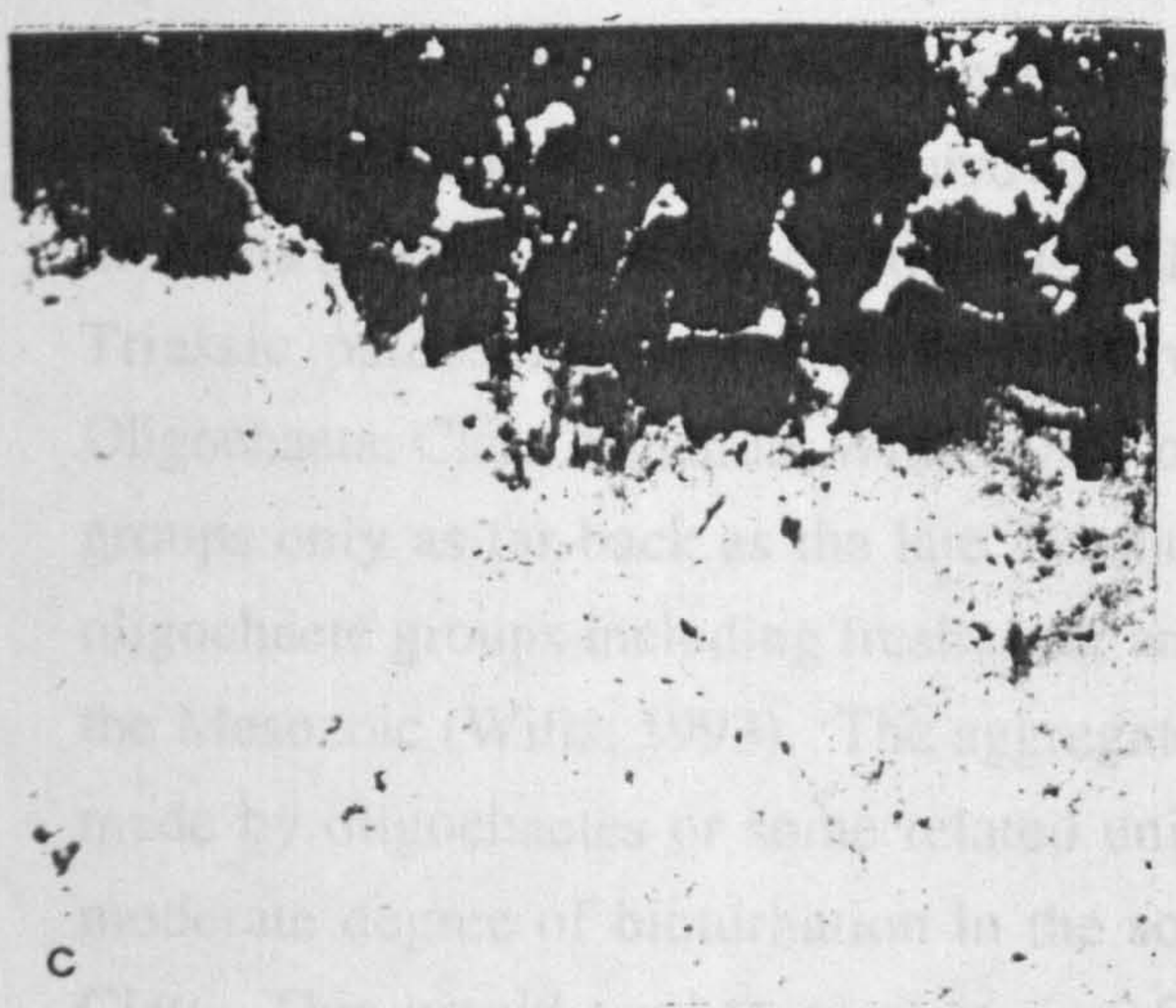
Pore space is more frequent where there are concentrations of unstructured organic matter and organans are frequently seen in the sections (Figs. 5.14a and 5.16a). Diffuse regions of ferruginised clay occur along some ped surfaces and in some cases skeleton carbonate clasts show strong micritic weathering rinds which are impregnated and coated with opaque granules of limonite. These rare ferrans suggest that the organic-rich clay horizon was at least periodically well-aerated.

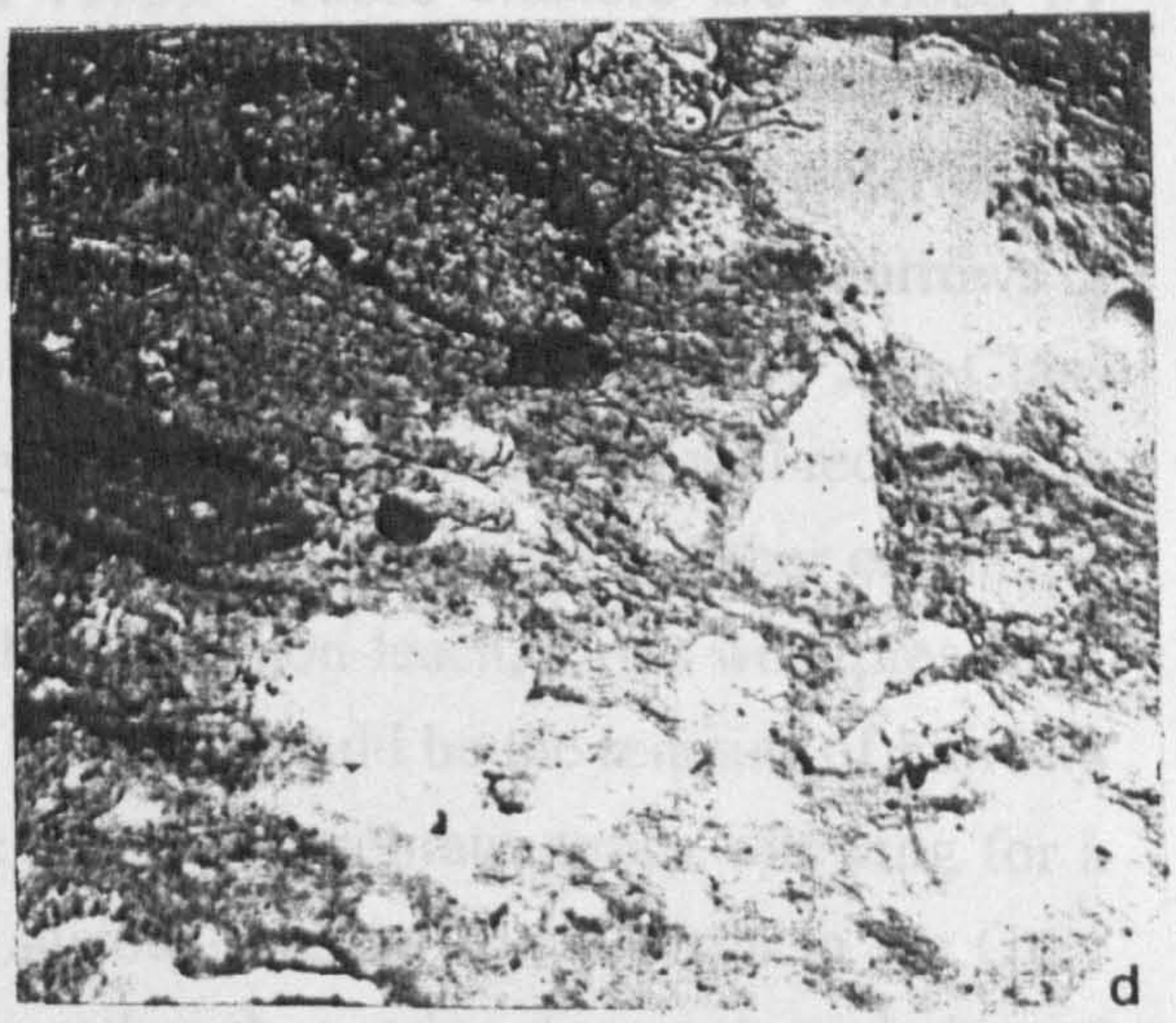
The most interesting features seen in thin section are thin vein-like regions of fine quartz sand. These tubular structures are best observed in section HQS12-4 and are approximately 1-2mm in width and several millimetres in length. They could be described as 'clastic dikes' (cf. Retallack, 1990) or illuviation skeletans. In detail, the dikes are filled with closely packed, sub-angular to sub-rounded grains of fine to medium sand (Figs. 5.14b and 5.16b). Over 70% of this sand is composed of quartz grains, although rare allochems also occur (Fig. 5.14b). These aggregates exhibit a coat and bridge structure of organic matter and clay, which are probably illuvial in origin. Pore space is limited and occurs as unconnected irregular pores between grains not bridged by the organan or argillan coatings.

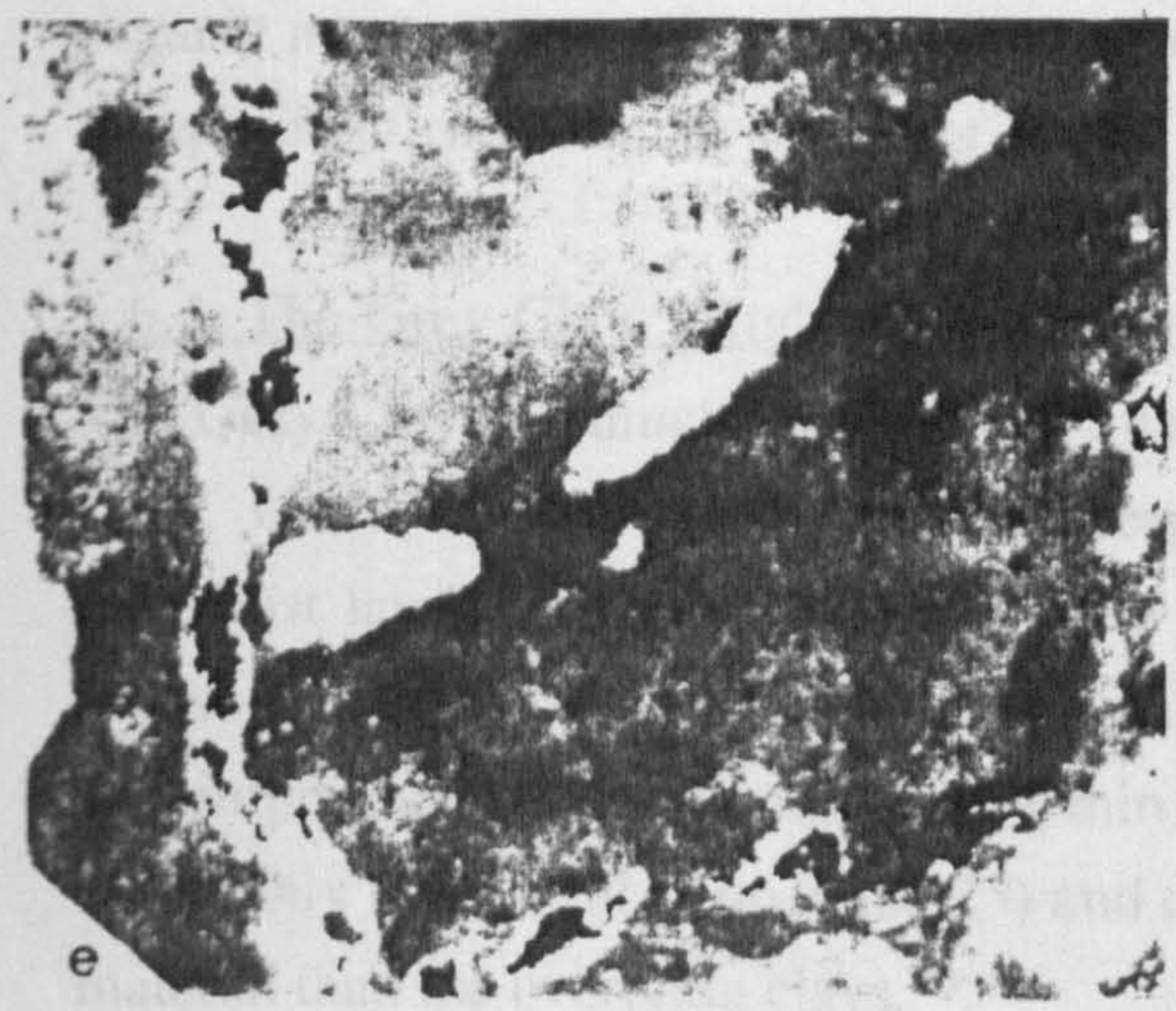
This almost classic example of coat and bridge structure is somewhat of an enigma in the usually clay-rich soil unit. Coat and bridge structures usually occur in sandy horizons where there has been deposition and concentration of clay and organic matter around the sand grains by illuviation. Organic coatings can form in sandy soils











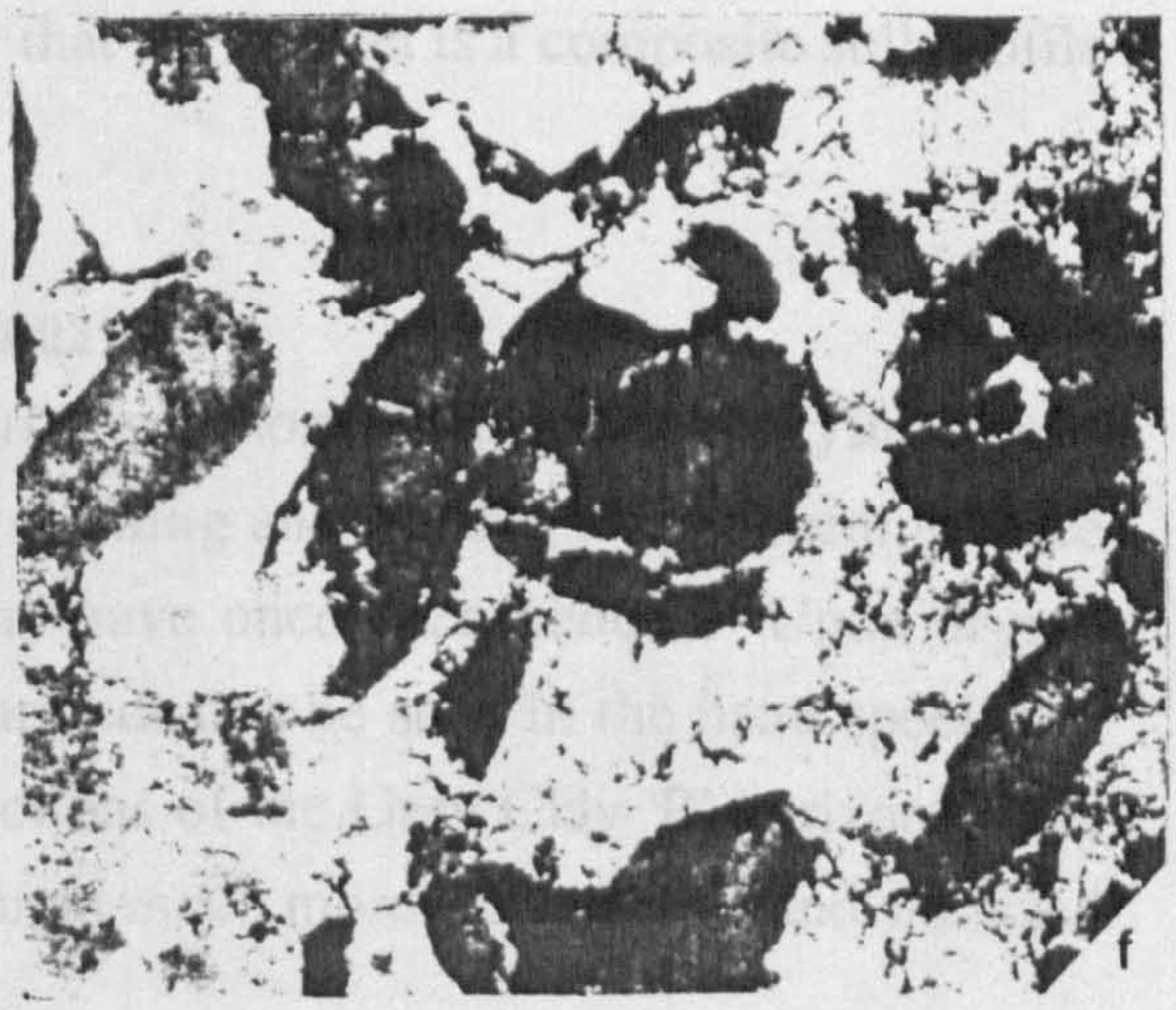


Figure 5.14. Photomicrographs of petrographic thin sections of the Grey Clay paleosol: (a) section HQS12-4, plane polarised light; (b) HQS12-4, plane polarised light; (c) HQS9-4, plane polarised light; (d) HQS11-5, plane polarised light; (e) HQS12-6, plane polarised light; (f) HQS12-6, cross polarised light. Field of view 2.3mm, except in (a) 0.9mm. All sections were photographed under blue-filtered light, except (a).

overlain by well-decomposed peat and these coatings exhibit a distinctive pattern of fine-scale cracks when viewed under high magnification (FitzPatrick, 1993). The organans seen upon the sand in HQS12-4 show this pattern. FitzPatrick (1993) describes similar organic/clay coatings upon sand grains and suggested that they form in the middle horizons of some Podsols. However, podzolisation usually occurs under highly acidic ground cover in cool and humid environments (e.g. coniferous or alpine forests, Retallack, 1990) and is accompanied by leaching and removal of clay to leave quartz-rich residual soils! Clearly, this is not the case with the Hornsleasow soils which are clay- and base-rich (section 5.10).

As the sandy aggregates occur in small vein-like structures they might be equivalent to the sand-grain clusters or 'pea-gravel' which is commonly observed in the topsoils of modern day soils (FitzPatrick, 1993). These clusters are formed by redistribution of the soil components by earthworm activity over a sustained period. Earthworms tend to line their burrows with sand to improve drainage in waterlogged horizons (Webster, 1965). Although Retallack (1976) has recognised similar burrows in Triassic paleosols, the terrestrial earthworms (Family Lumbriculidae, Sub-Class Oligochaeta, Class Annelida; Wills, 1993) only have a relatively recent fossil record, with groups only as far back as the late Tertiary (Pliocene; Wills, 1993). Other enigmatic oligochaete groups including freshwater and brackish, non-leach, forms were present in the Mesozoic (Wills, 1993). The aggregates of quartz could be the remains of burrows made by oligochaetes or some related unidentified invertebrate group, allowing for a moderate degree of bioturbation in the soil surrounding the 'Oe' horizon in the Grey Clay. This would tend to suggest a surface rather than subsurface accumulation of organic matter, and lends support to the view that the section is a composite soil profile composed of two or more soil units.

5.6.4. The Grey Clay '2Btg' and 'Btkg' horizons

The Grey Clay is a much less massive appearing soil horizon than the overlying Green Clay. Lines of skeleton clasts, diffuse ferric mottling and organic matter seem to trace out ghost lamination, and the Grey Clay may have once been bedded. Upon drying coarse angular blocky and finer-scale laminar peds can be seen in the hand specimen. The impregnation and preparation of thin sections of the Grey Clay 'B' horizons were reasonably successful (Appendix C5) and contain much more detrital sand and silt sized material than the overlying clays.

The structure of the 'B' horizon ranges from massive continuous soil phase (Figs. 5.14c and 5.16c) with few complete pore spaces, to a irregular blocky and laminar structures (Figs. 5.9, 5.14d and 5.16d). This composite and variable structure may indicate seasonal variation in conditions such as periodic waterlogging and desiccation,

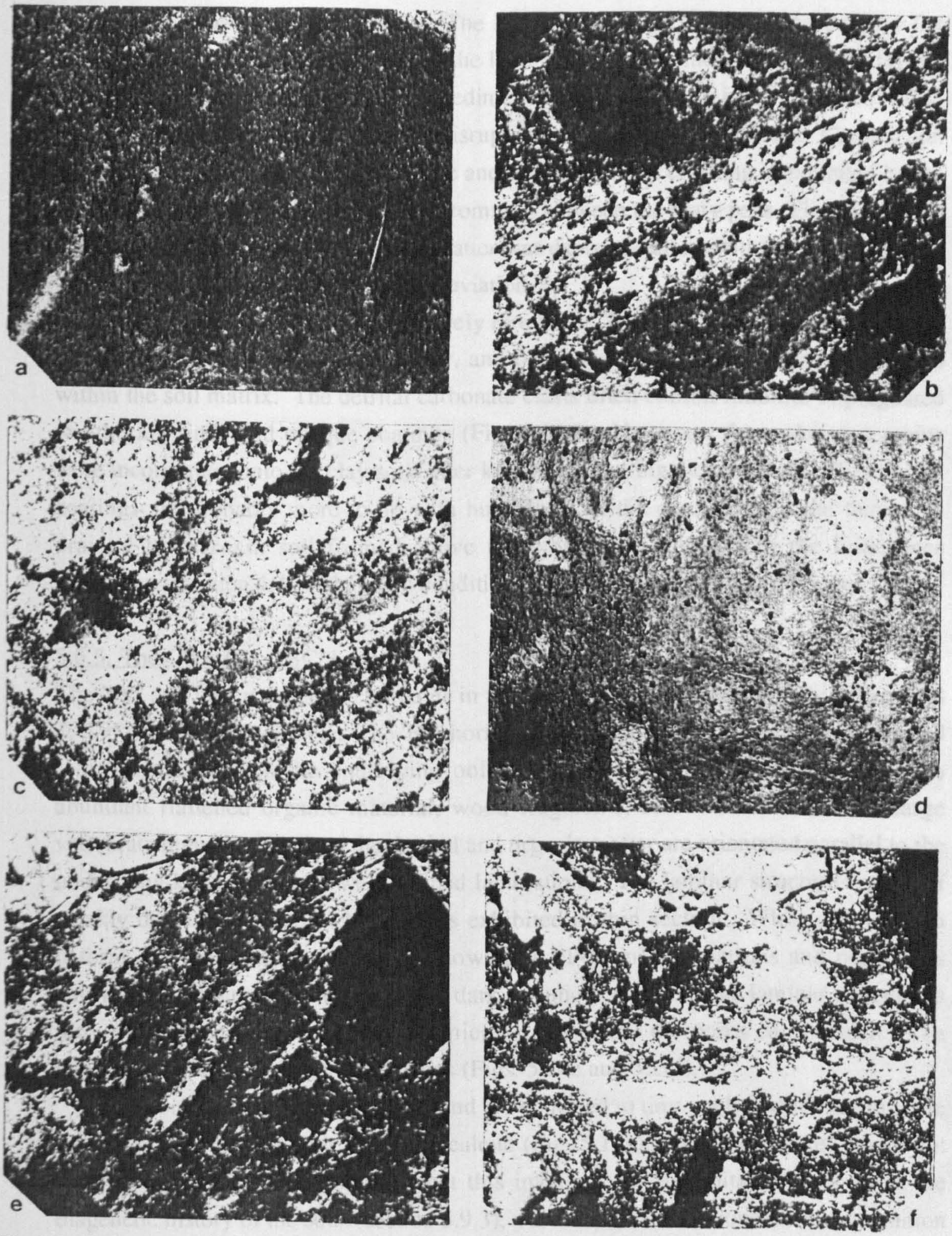


Figure 5.15. Photomicrographs of petrographic thin sections of the Green Clay paleosol: (a) section HQS12-1, plane polarised light; (b) HQS9-1, cross polarised light; (c) HQS12-1, plane polarised light; (d) HQS11-2, plane polarised light; (e) HQS11-2, cross polarised light; (f) HQS11-2, cross polarised light. Field of view 2.3mm, except in (a-c, e) 0.9mm. All sections were photographed under blue-filtered light, except (a).

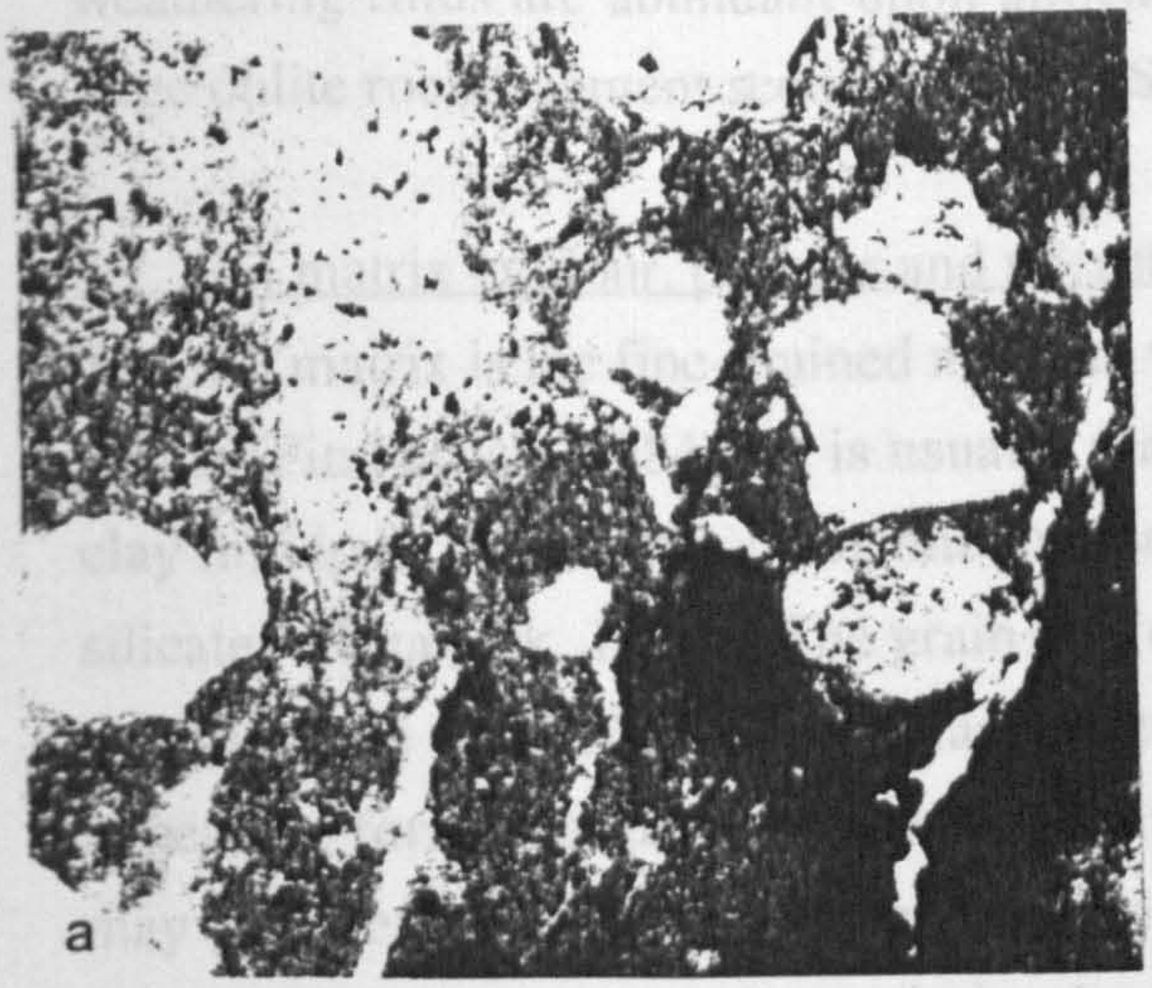
and the preservation of discrete incomplete bifurcating shrinkage cracks within massive soil regions support this hypothesis. The laminar structure is interesting because the platy peds are orientated horizontal to the land surface. The most likely origin for the platy peds and laminar structure are as sedimentary laminations or bedding planes within the clay, which have only begun to be disrupted by soil-forming processes. Compaction is likely to enhance the laminar structure and some degree of crushing is exhibited by the lack of interpedal pores. Organans are common between the platy peds (Figs. 5.14d and 5.16d) and it is likely that the organic matter was deposited along with the clay, although some may have been incorporated by illuviation.

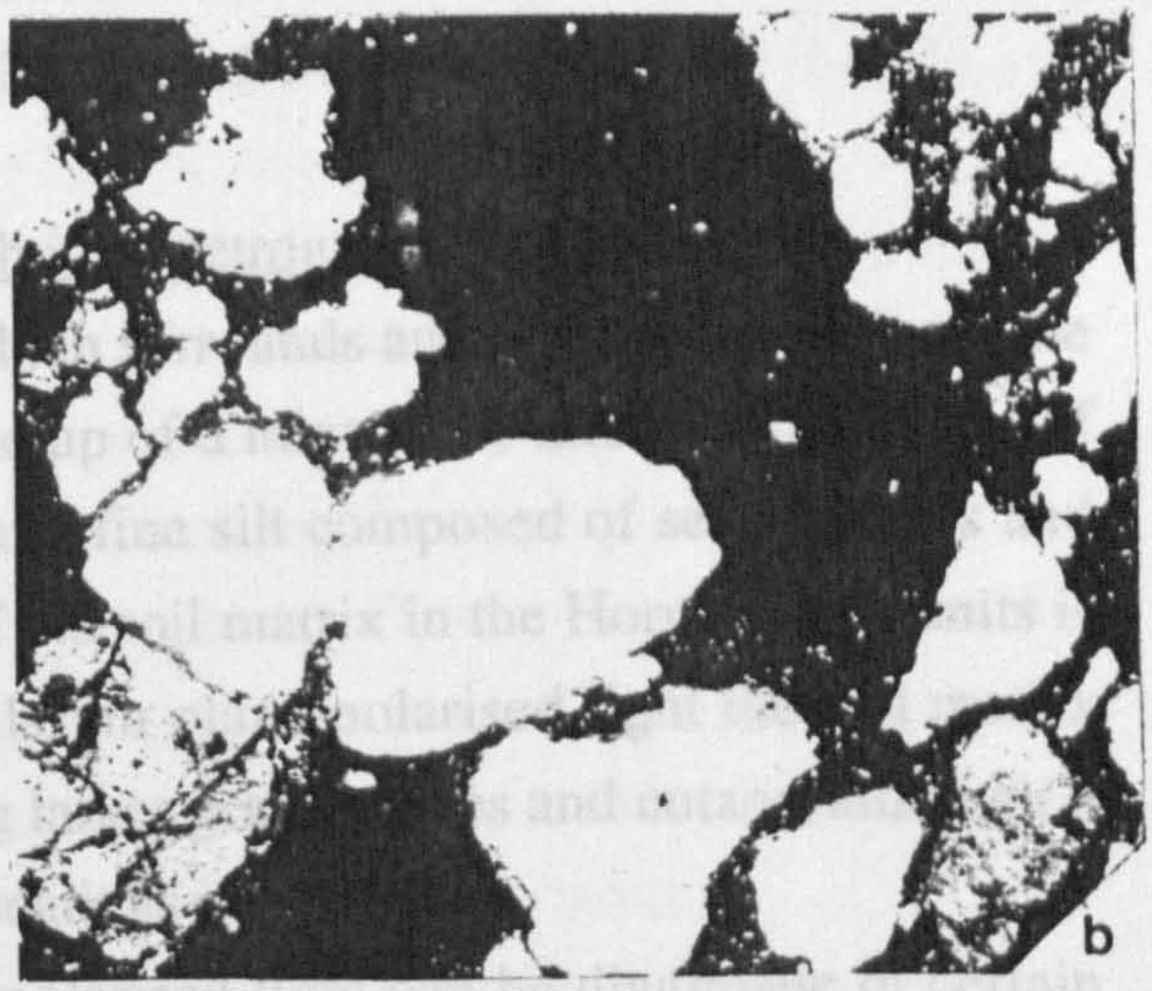
Other types of cutans are relatively rare in the 'B' horizon, although ferruginised clay is associated with organic matter, and organans and ferrans are not uncommon within the soil matrix. The detrital carbonate clasts often exhibit limonite-impregnated weathering rinds and opaque coatings (Fig. 5.14d). However, these skeleton grains were incorporated into the clay soils after karstic weathering (section 4.4). Their ferran coatings are probably more to do with humid and acidic conditions in the thin well-drained *terra rossa* soil formed above the palaeokarst and within the limestones themselves (section 4.4), rather than conditions within the Grey Clay soil horizon.

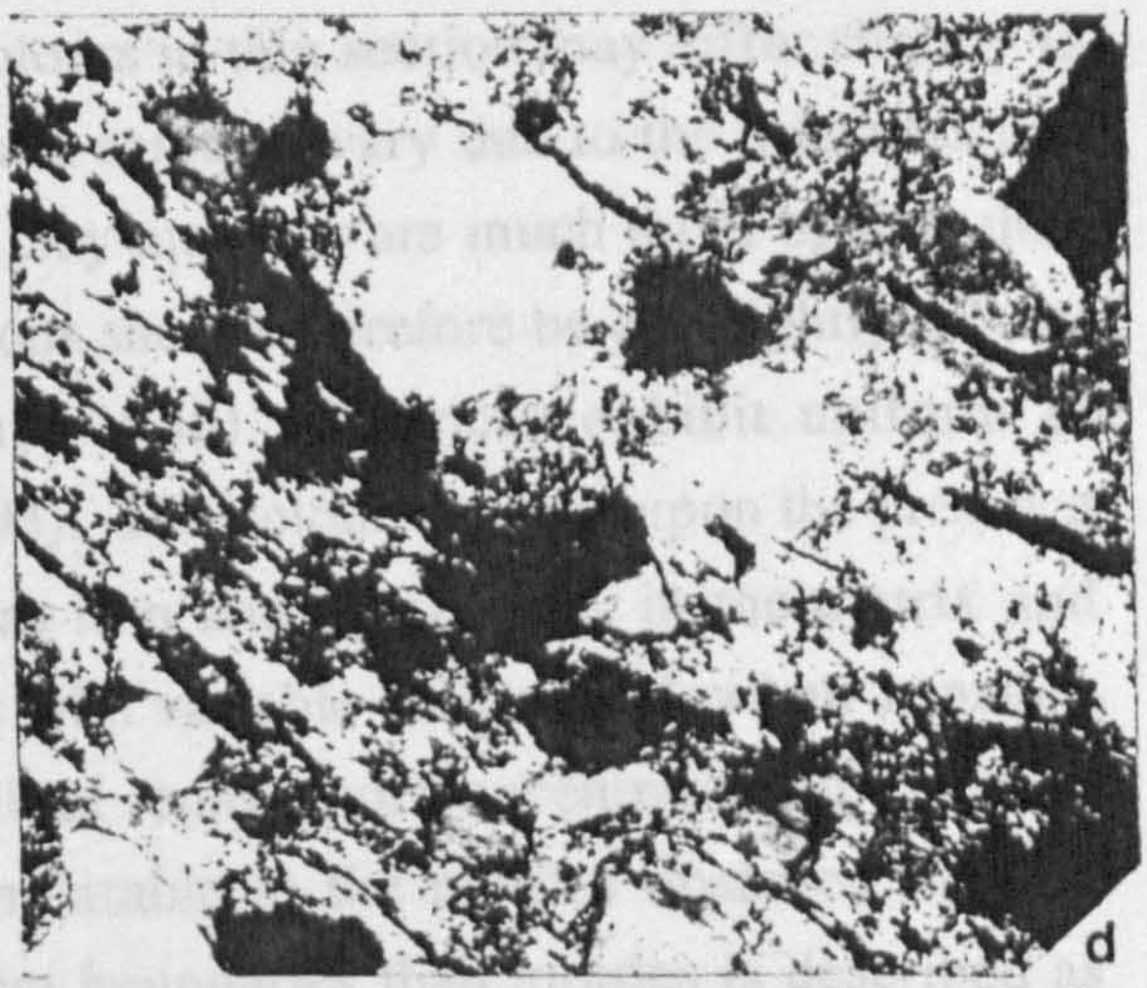
5.6.5. The Grey Clay 'BC' horizon

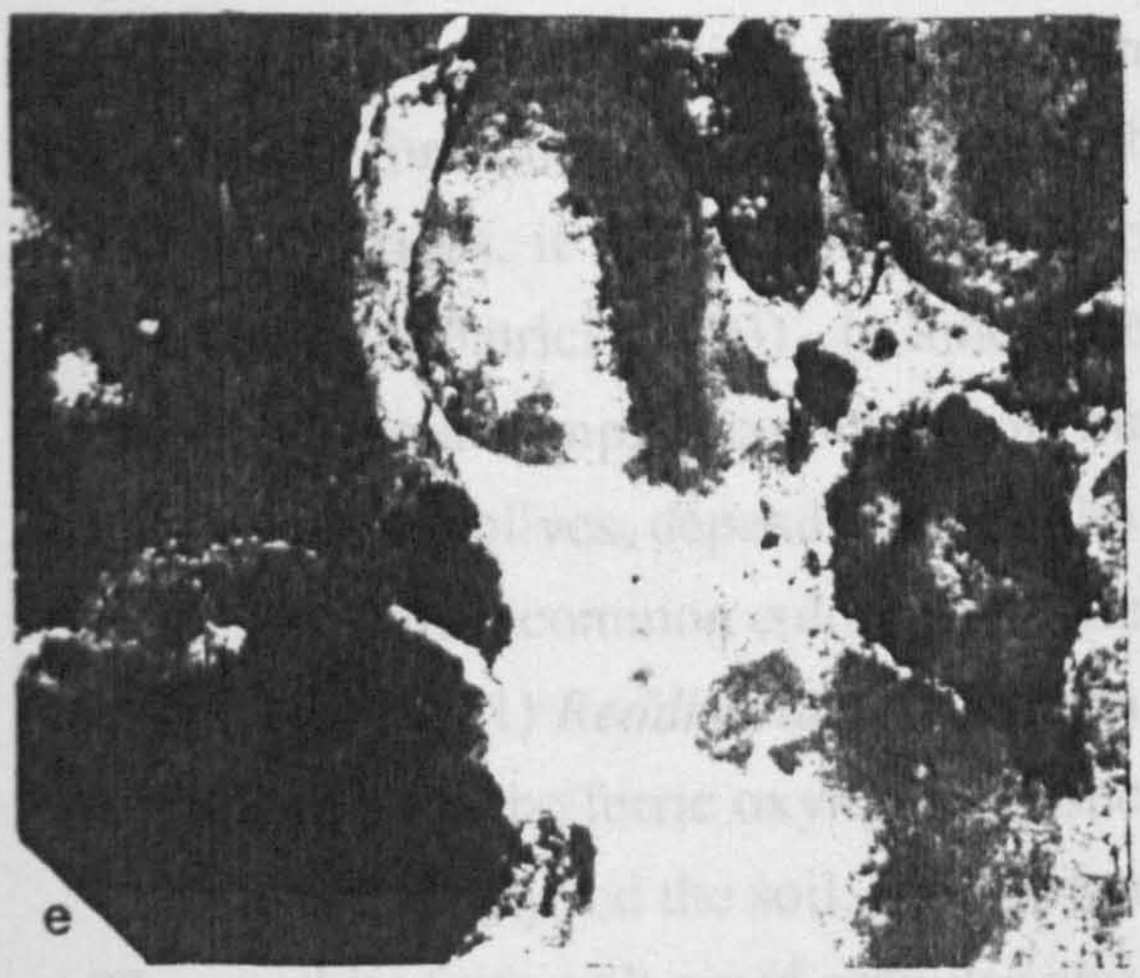
The 'BC' and underlying 'Cro' horizons in the soil profiles are extremely thin (Figs. 5.8 & 5.9) and laterally variable. The 'BC' horizon contains almost 50% skeleton grains and rock fragments from the underlying oolite (Figs. 5.14f and 5.16e,f,) and has also abundant flattened organic material, wood fragments, carbonate nodules and large vertebrate remains. Much of the detrital and organic matter are orientated parallel to the underlying karst surface in well-defined laminations. The laminar structure typical of weakly developed paleosol horizons is exhibited in thin section. When examined in natural daylight, section HQS12-6 shows thin (1-2mm) laminations and platy peds defined by thin sinuous pore space and dark organic ped skins. The laminar structure is also visible when viewed under the microscope, where the pores can be seen to be continuous with occasional bifurcations (Figs. 5.14e and 5.16f).

Interpedal pores are quite rare and are confined to tiny vughs, which are in some cases partially infilled with a micritic calcite (Fig. 5.14e). That these pores where not compacted during burial suggests that this infilling was precipitated early on in the diagenetic history of the soils (section 5.9.3). As in the other horizons the most common types of cutans observed were organans and some diffuse weakly developed ferrans. Mangans occur in this horizon upon the larger oolite fragments and some of the large bones and these associated with the initial change in conditions from karstic weathering to waterlogged clay-rich soils (section 4.5). Ferric oxyhydrite coatings and micritic









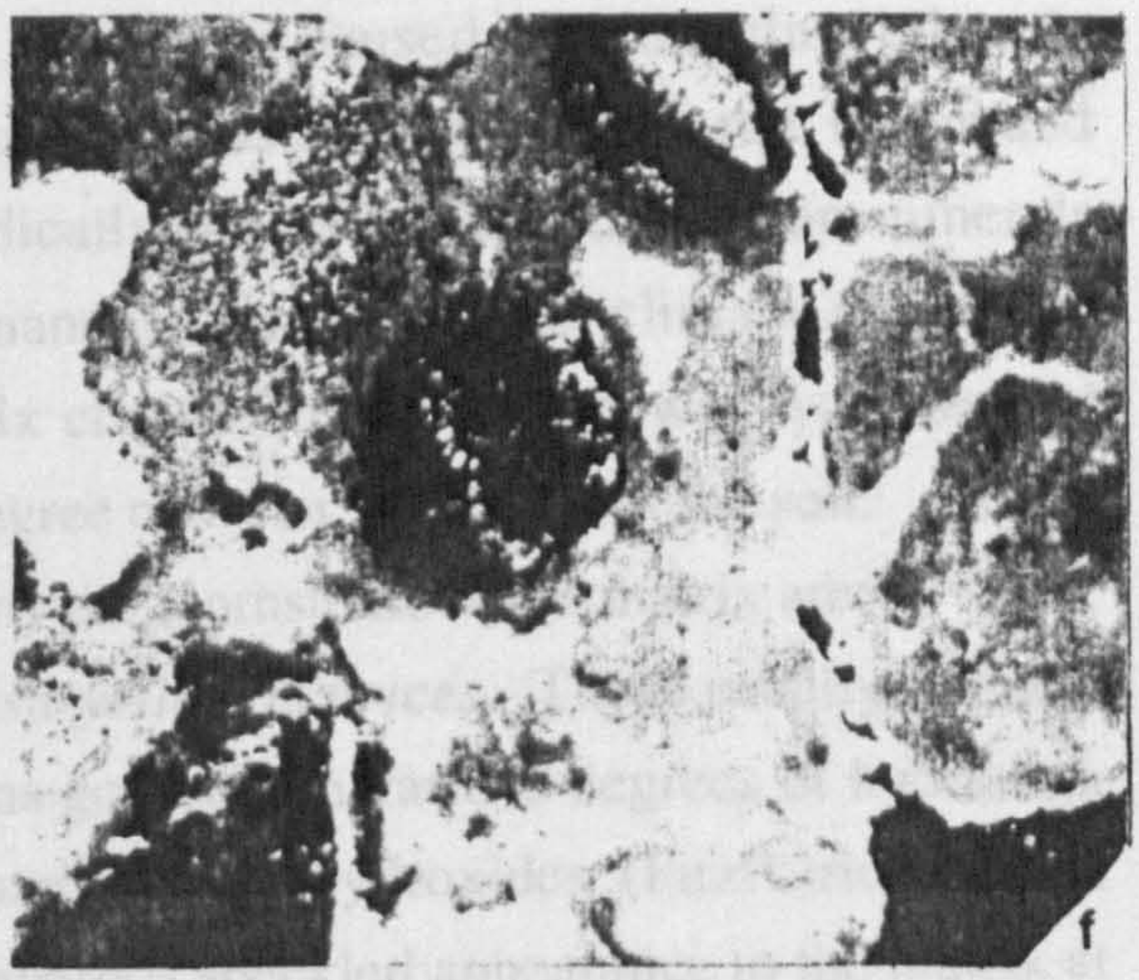


Figure 5.16. Photomicrographs of petrographic thin sections of the Grey Clay paleosol: (a) section HQS12-4, plane polarised light; (b) HQS12-4, cross polarised light; (c) HQS12-5, plane polarised light; (d) HQS11-5, cross polarised light; (e) HQS12-6, plane polarised light; (f) HQS12-6, cross polarised light. Field of view 2.3mm, except in (c,d) 0.9mm. All sections were photographed under blue-filtered light.

weathering rinds are abundant upon allochems, and these were also observed within a large oolite rock fragment sectioned in HQS12-6 (Fig. 5.14f).

5.7. Soil matrix - colour, patterns and microfabric in petrographic thin section

The soil matrix is the fine-grained material which surrounds and completely encloses the grains (FitzPatrick, 1984). It is usually made up of a mixture of microscopic grains of clay minerals, disseminated organic matter and fine silt composed of sesquioxides and silicates (Retallack, 1990). The grain size of the soil matrix in the Hornsleasow units is largely made up of fine silt and clay minerals. In plane-polarised light the soil matrix appears to form a continuous phase enclosing larger grains, pores and cutans, although it may be differentiated into argillans and clay coatings in some soils.

The colour of a soil matrix in plane-polarised light can be diagnostic of certain conditions at soil formation or alteration. Colours in thin section may differ slightly to those seen in hand specimens (section 5.4) and can also vary due to the thickness and impregnation of the thin section. As a rule, clay matrices are much more opaque than common sedimentary minerals and soil sections should therefore be cut slightly thinner than standard petrographic slides (Murphy, 1986). Matrices exhibit uniform to complicated colour patterns (FitzPatrick, 1993). The colour depends upon the presence of different colour-bearing elements, such as iron and manganese, in the matrix and whether organic matter is present. Matrices with variable colours and colour zonation have usually undergone some form of localised removal or concentration of material. The colour variations seen are broadly comparable to the mottles observed in hand specimen. Colour variation with less distinct boundaries than mottles is described as being 'marbled'. Mottling and marbling is usually caused by changing climatic conditions, for example, a red matrix may become diffusely marbled with red, brown and yellow matrices, if the soil becomes periodically waterlogged and the iron minerals hydrated (FitzPatrick, 1993). In time, a permanent change to wetter climatic conditions could lead to a complex coloured soil matrix composed of reds, browns, yellows and even greys and olives, depending upon the degree of water saturation in the soil.

The most common colours exhibited by the Hornsleasow soil matrix are:

(1) *Reddish-brown to brownish-yellow matrices.* These usually indicate the presence of the ferric oxyhydrites, such as goethite, in various degrees of hydration. The more strongly red the soil, the less hydrated the iron hydroxides (FitzPatrick, 1993). The goethite forms silt-sized granules and imparts a speckled appearance to the matrix at moderate and high magnifications. These matrix colours are common in tropical and sub-tropical soils, and are probably caused by the localised segregation and concentration of iron-bearing minerals. An 'ochreous' brownish yellow mottled matrix may also indicate the presence of finely disseminated weathering product, such as limonite. This hydrated

mineral is the principal iron oxide in wet soils, where it forms the characteristic yellow mottles in the grey matrix. The mottles tend to follow the pore system in the soil and form due to fluctuating oxygenation in seasonally wet soils (FitzPatrick, 1993).

(2) *Red to reddish-yellow matrices.* This colour is relatively rare in soils and usually indicates the presence of anhydrous iron oxides in the form of haematite. This matrix is not observed in the clay soil units at Hornsleasow, but is present in the *terra-rossa* soil and karstified limestone directly underlying the clays (section 4.4.2).

(3) *Dark brown and brown matrices.* These colours are usually associated with an abundance of colloidal organic matter or amorphous limonite and other ferrihydrite minerals (FitzPatrick, 1993). In cross-polars these matrices are isotropic. These matrices are commonly associated with organic-rich, wet horizons.

(4) *Olive-green, olive-grey and grey matrices.* These drab colours are associated with reducing conditions and are caused by the presence of ferrous iron compounds. They are most abundant in soil horizons which are seasonally or permanently waterlogged (FitzPatrick, 1993).

(5) *Pale brown and pale-grey matrices.* These are matrices which have lost coloured substances and minerals during leaching. Pale matrix colours may also occur in sections ground too thinly.

(6) *Opaque matrices.* These are similar to the red haematitic matrices described above. They consist of haematite and/or pyrolusite coatings and can be studied in reflected light to differentiate between the two minerals (FitzPatrick, 1993). Opaque matrices are associated with deep, humid weathering and occur in indurated subsurface horizons such as laterites and bog-iron ores (Retallack, 1990), again it is the karstic limestone and *terra-rossa* soil which show opaque and red matrices, rather than the overlying clays. Poorly impregnated and coarsely ground sections may appear opaque and unstructured carbonised organic matter is also opaque.

The microfabric of the clayey soil matrix can also be studied in cross-polarised light. Translucent clay matrices either appear completely black or optically 'isotropic' when viewed in cross-polars or they show patterns of different interference colours or 'birefringence'. The latter are termed 'anisotropic' or 'bright clay' matrices (FitzPatrick, 1984, 1993, Retallack, 1990). Most clay minerals are optically anisotropic, but as they occur naturally in fine particulate form, it is their organisation within the matrix which governs whether the fabric will show interference colours. If the clay particles are aligned and closely packed, the resulting microfabric will show bright interference colours in cross-polarised light, whilst randomly orientated particles give the impression of an isotropic matrix. Few soil matrices are completely isotropic although they may appear uniformly dull or featureless, or have regions of sombre cloudiness (Retallack, 1990). It is thought that dull fabrics are caused by the drying of colloidal material such as

clay mixed with organic matter or opaques, and these are common in the clayey parts of waterlogged swampy soils underlying peats (Retallack, 1990).

The single most important factor in the formation of matrix anisotropism is pressure (Green-Kelley & Mackney, 1970). This is especially true in sediments and soils, where shear stresses caused by differential expansion and contraction of clayey matrix during alternating wetting and drying episodes, produce lines of aligned particles called 'slickenslides'. Other factors which may cause alignment of clay particles within soils are illuviation into cracks and pore space, and the surface weathering of individual peds. The destruction of original sedimentary layering and the formation of argillans is expressed on the microscopic level as the development of the bright clay or anisotropic microfabric exhibited in soil matrix (Retallack, 1990).

In this account I followed FitzPatrick's (1984, 1993) terminology in all descriptions and classification of matrix micromorphology, as I believe it is the easiest to use, the simplest to define and probably the most diagnostic. There are a number of other microfabric classifications for soils, including the one favoured by the *Handbook for Soil Thin Section Description* (Bullock *et al.*, 1985) and that proposed by Brewer (1976) in his *Fabric and Mineral Analysis of Soils*. Brewer (1976) was the first to describe and classify soil matrix microfabric but his scheme includes a complicated system of hierarchical Latinate jargon which is unwieldy to use in practice. For instance, Brewer terms soil matrix 'plasma', the microfabric as 'plasmic fabric' and features such as streaky bright clay as 'mosepic plasmic fabric' (Brewer, 1976). Bullock *et al.* (1985) have attempted to combine Brewer's terminology with the simple terms suggested by FitzPatrick (1984) and consequently their scheme suffers from too much jargon, is not consistent and contains some grossly inaccurate descriptions.

The scheme outlined in FitzPatrick (1984, 1993) is not perfect, as it does not contain as many groups as the other two schemes and in common with these, it can suffer from an over-splitting of groups. However, the terms used are simple and diagnostic of the particular feature they describe. His scheme also has the advantage of being internally consistent. Retallack (1990) calls matrix anisotropy, 'bright clay microfabric', and the terms are fairly interchangeable in the following descriptions.

The four major descriptive classes of anisotropy in soil matrices are defined by FitzPatrick (1993) as anisotropic flecks, anisotropic zones, anisotropic aureoles and anisotropic lines. The different patterns of bright clay microfabric reflect different conditions of formation and these are:

(1) *Anisotropic Flecks*. These are the discrete small areas of optically anisotropic clay particles which are closely packed and aligned. They are also known as 'domains' (FitzPatrick, 1993). Anisotropic flecks have diffuse boundaries, low order interference colours, and show no appreciable difference in relief than the surrounding

matrix in plane polarised light, which helps to differentiate them from similarly-sized mica flakes. Anisotropic flecks are considered to be the earliest sign of mineral particle alignment and are termed 'weakly anisotropic matrices' (FitzPatrick, 1993). They occur within immature and/or gleyed soils, where original bedding is preserved. Alignment of single domains as a result of disturbance by bioturbation or soil development produces the anisotropic zones, aureoles and in some instances, anisotropic lines.

The interference colours and size of anisotropic flecks are variable and can be diagnostic of a particular clay mineralogy. For example, kaolinite usually displays first order greys and white birefringence, but if iron is present then the flecks are orange or red in cross polarised light. In deeply weathered and oxidised soils, the flecks will be short and narrow (10x1.0µm), whilst in gleyed soils they are long and thick (50-100x10µm), indicating a composition of kaolinite or even micaceous minerals (FitzPatrick, 1993). Swelling clays such as smectite tend to have short and narrow domains (10-50x1-5µm) which frequently are aligned in zones, aureoles or lines.

Anisotropic flecks may also be randomly organised or aligned to give an impression of overall orientation (FitzPatrick, 1993). In clay sediments unaffected by churning soil possesses, anisotropic flecks are horizontally orientated with sedimentary lamination or bedding.

(2) *Anisotropic Zones.* These occur when anisotropic flecks become aligned and converge to form a zone of optical anisotropism with uniform optical properties which run through the soil matrix. Anisotropic zones are the main features to be described in all the systems of soil microfabric classification (cf. Brewer, 1970 'sepic plasmic microfabric'; Bullock *et al.* 1985 'birefringence streaks'; Retallack, 1990 'incipient, streaky and fibrous bright clay') and are common within the Hornsleasow paleosols (Appendix C5).

There are many different patterns of anisotropic zoning (cf. Brewer, 1976), but these can be simplified into five categories, in order of increasing alignment: striated; fibrous; continuous; reticulate; and variegated anisotropic zones (FitzPatrick, 1993). (i) Striated anisotropic zones are characterised by displaying an internal striated or streaky pattern of loosely packed aligned anisotropic flecks. They have fairly diffuse boundaries and are not normally orientated to any surface or one another, although they may form argillans and can be arranged in normal or oblique orientation to each other (FitzPatrick, 1993). (ii) Fibrous zones are composed of long thin anisotropic flecks or clusters of domains which are generally aligned in a parallel arrangement. They appear as well-defined streaks of bright clay, and the alignment of the clay particles can generally be seen as uniform extinction as the stage of the microscope is rotated. They may conform to the surface of a soil structure and have a variety of different orientations (FitzPatrick, 1993), although they usually occur as multiply parallel aligned clusters. (iii) Continuous

anisotropic zones appear as diffuse streaks which run through the matrix and occur where numerous domains have coalesced. They often conform to and may define a surface of a ped or pore and are common examples of argillans, or they can occur as single or multiple zones. Multiple continuous zones tend to be organised in some way, either developing a criss-cross, radial or concentric pattern, or being aligned parallel to one another (FitzPatrick, 1993). (iv) Reticulate zones of anisotropy are composed of a criss-cross network arrangement composed of fine intersecting anisotropic flecks. This microfabric is best viewed under circularly polarised light (FitzPatrick, 1993). Reticulate zones are also described as lattice, net or trellis-like bright clay microfabrics (Retallack, 1990). Reticulate zones are seen within soil matrices which have undergone substantial weathering. In very well developed and well drained soils they may exhibit a woven-like arrangement of anisotropic flecks (Retallack, 1990). Although, they usually form in very old soils, they can also form rapidly within soils which contain expandable clays. (v) Variegated anisotropic zones are similar to reticulate zones, but are less clearly defined. They are irregular in shape and are best distinguished from reticulate zones in circularly polarised light (FitzPatrick, 1993). These zones are formed in subsurface clay-rich 'Bt' horizons, where the matrix is under a moderate confining pressure (Retallack, 1990).

(3) *Anisotropic aureoles.* These are concentrically arranged regions of aligned anisotropic flecks which tend to form around pores, skeleton grains and rock fragments within the soil matrix. They are variable in thickness, showing diffuse outer contacts which tend to decrease in brightness away from the grain or pore, and they may partially or completely surround the object in the matrix. Internally, anisotropic aureoles are composed of striated, fibrous or continuous orientated anisotropic flecks, depending upon the degree of 'kneading' of the matrix by bioturbation, and the size and distribution of particles (FitzPatrick, 1993). Essentially, the rule is that the larger the grain-size and the more loosely packed the object is within the matrix, the larger and better defined the surrounding aureole (Smart & Tovey, 1981).

(4) *Anisotropic lines.* These are common forms of stress argillans or slickenslides (section 5.5), as they can conform to ped surfaces. However, they can also cut across earlier fabrics and are generally regarded as thin layers of sheared and pressure-aligned clay particles (FitzPatrick, 1993). They are usually only a few microns in thickness, depending on the thickness of the constituent clay particles. When anisotropic lines occur within the matrix, but are not argillans, they are regarded as slickenslides where the two opposing surfaces of the shear plane have not separated or have been rejoined (Smart & Tovey, 1981).

The matrix of a soil is not static and the fabric can be variable over the space of one thin section (FitzPatrick, 1993). The development of anisotropic microfabric is due a combination of the time available for soil formation and the intensity of the soil forming

processes. Fabrics are formed and modified by changing physical, biological and chemical conditions within the soil. For example the stresses brought about by shrinkage and expansion in seasonally waterlogged clay soils will cause the deformation and re-establishment of anisotropic patterns in the matrix. Shrinkage may cause disruption and shearing of established anisotropic zones, whilst wetting may cause flow patterns such as anisotropic aureoles. Bioturbation and soil ingestion also modifies soil matrices, and finally, chemical weathering will change the mineralogy and physical features of the individual domains (FitzPatrick, 1993). Anisotropic matrices may also form in sediments unaltered by soil forming processes, for example laminations within marine clays may show orientated bright clay fabrics and anisotropic aureoles can form around grains during compaction (Retallack, 1990). However, the presence of anisotropic zones usually indicates formation in a soil environment.

5.7.1. The Green Clay 'Aac' horizon

This is characterised by strong iron-stained mottling and ferruginised nodules. In thin section, the matrix of the soils also exhibit a strong colour mottling and this is seen in both plane and cross polarised light. The overall colour of the matrix under plane polars is a pale yellowish brown (Fig. 5.13a,b). The matrix also appears to be speckled, and this is probably due to silt-sized sesquioxide and organic particles trapped in the matrix (Fig. 5.13a). Some larger 'silt' particles are actually contamination in the slides, these are the remains of the carborundum abrasive used to polish the sections and are easily distinguishable. Within the mottles the matrix is clearly darker and more reddish brown in colour. This is best seen in section HQS12-1 (Fig. 5.15a), where the boundaries of the mottles are reasonably defined. The mottling is associated in this section with organic matter and disseminated limonite. Mottles are also seen in close proximity to weathered carbonate allochems, rootlets and conforming to some ped surfaces (Fig. 5.13b).

Under cross-polars the matrix displays a moderate to strong amount of anisotropic microfabric (Fig. 5.15b,c). Ped surfaces and pores are often defined by bright fibrous anisotropic zones (Fig. 5.15b), with strongly aligned clay particle domains, whilst the interior of some peds show streaky birefringence (Fig. 5.15c) and randomly orientated anisotropic flecks. The striated and fibrous zones range from 1st order yellows to greys in unmottled regions, to bright orange and reds in the mottles. They become paler towards the edges of the slide and around large pores, where the section has been ground too thinly. The clasts within the matrix show diffuse, fairly large anisotropic aureoles (Fig. 5.15b), and these sometimes coalesce and join up into fibrous zones where the grains are closely packed. The aureoles show 1st order yellows and oranges. Section HQS12-1 exhibits anisotropic lines or slickenslides. These thin

regions of anisotropism appear to cut across earlier microfabrics and are commonly 1st order whites and greys (Fig. 5.15c).

The brownish-yellow matrix colour, the reddish-brown mottles and the anisotropism exhibited by the 'Ae' horizon of the Green Clay are in keeping with it being a fairly well-aerated and moderately mature soil horizon. The presence of limonite-stained mottles suggests formation in a seasonal climate.

5.7.2. The Green Clay 'Btg' horizon

The 'Btg' is strongly green coloured in hand specimen, however in thin section the matrix appears a pale buff-brown (Fig. 5.15d) or a brownish khaki (Fig. 5.13e,f). Warm or faint orange-brown mottles and marbled colour variation do occur also in some of the specimens (Fig. 5.13d), but they are not usually as well developed as those in the 'Ae' horizon sections. The orange-brown mottles are associated with rootlets (Fig. 5.13f) and some opaques, and they are also present around weathered micritised carbonate allochems. The matrix appears to be stained a deep brown in some regions were there is abundant disseminated carbonaceous matter (Fig. 5.15d).

Under cross-polars the clay matrices are moderately anisotropic, although the degree of anisotropism varies over the site and horizon (Fig. 5.15e,f). Most of the matrix exhibits bright 1st order colours, with greys, yellows and pale oranges being common in the normal matrix. Mottles are more strongly birefringent and exhibit 1st order reds and oranges, indicating a localised abundance of ferric iron in these regions. The matrix shows some well developed fibrous zones which run through the sections (Fig. 5.15e) and these may conform to surfaces of the blocky ped structure. In one or two sections, continuous anisotropic zones are also present indicating prolonged soil development in this horizon. The interior of the peds are usually strongly flecked and some show streaks of striated anisotropic zones running through them (Fig. 5.15e,f). Many of the quartz sand grains and some of the allochems show moderately well formed bright clay aureoles (Fig. 5.15e). Some of the latter exhibit red and orange interference colours indicating the diffusion into the matrix of iron oxidised during dissolution of the ferroan carbonates. A few organic particles show this effect too. The section HQS11-2 shows beautifully developed anisotropic lines which cross through and shear earlier microfabric (Fig. 5.15f).

The microfabrics and matrix colours exhibited by the 'Btg' horizon of the Green Clay suggest a moderate amount of weathering and soil structure formation. The speckled khaki-brown and brown matrix suggests that the horizon may have been periodically waterlogged, but the presence of oxidised iron in the clay suggests a fluctuating moisture level and fairly well aerated soil.

5.7.3. The Grey Clay 'Oc' and 'E/B' horizons

The two sections removed from around the organic rich layer at the top of the Grey Clay contain abundant colloidal and larger patches of carbonaceous organic matter (Figs. 5.14a and 5.16a). The fine disseminated material has imparted a speckled dark brown colour to the matrix, almost becoming opaque in regions close to large fragments of organic matter. In some areas close to the organics the matrix exhibits a deep reddish-brown colour (Fig. 5.14a). As much of the carbonised debris are associated with limonitic stains and ferrihydrite replacement, this colour is probably associated with iron oxides being leached out of the decayed plant matter. This strong iron content of the matrix around the organics is also shown in cross polars, where the clays show deep reds and orange birefringence.

The matrix anisotropism is not particularly well developed in the organic rich horizon and in some places the matrix appears dull and almost isotropic in cross polars. Much of the matrix shows regions of diffuse fine anisotropic flecks, which are commonly 1st order yellows and greys. Other microfabric features are rare, with weakly developed thin bright clay aureoles being observed around closely-packed ferroan carbonate allochems, which show pervasive micritic weathering rinds. In section HQS12-4, a few wispy fibrous anisotropic zones where noticed in the more clay-rich regions and the interior of 'peds' show the weak development of striated microfabric and alignment of anisotropic domains. The clay matrix which surrounds the quartz sand in the 'pea-sand' aggregates is saturated with opaque carbonaceous organic matter, it shows no birefringence in cross polars and appears almost black (Fig. 5.16b).

The strong dark brown colour is largely a function of the high organic concentration in this horizon. The weak or absent anisotropism is also associated with the high content of amorphous organic matter disseminated through the clay matrix and the associated finely crystalline ferrihydrite granules (FitzPatrick, 1993). The strong colours have masked any information upon aeration and moisture content of this horizon.

5.7.4. The Grey Clay '2Btg' and 'Btkg' horizons

Of the three sections taken from the 'B' horizon, HQS9-4 may have more in common with the overlying 'Oe' horizon than was previously suspected. The 'O' horizon is not present directly beneath the Grey/Green boundary in the block section S9 (Fig. 5.7), although there is abundant carbonaceous material further down. However, HQS9-4 is packed with organic matter and consequently shows a speckled buff-brown colour with deep brown mottles close to plant material in plane polarised light (Fig. 5.14c). In contrast, the other sections HQS11-5 and HQS12-5 have a pale yellowish brown to khaki coloured matrix (Fig. 5.14d), although they also show deep orange-brown mottles

associated with plant matter (Fig. 5.16c). The slide HQS12-5 is also heavily contaminated with carborundum powder (blue-tinged, high relief silt-sized particles; Fig. 5.16c).

In cross polarised light, the matrices show a weak to moderate anisotropic microfabric, which is characterised by diffuse areas of aligned flecked anisotropy which may coalesce into streaky or striated anisotropic zones (Fig. 5.16d). Fibrous anisotropic zones are associated with the laminar peds developed in the 'Btg' section HQS11-5 and in a few cases the areas are substantial enough to be considered as continuous zones of bright clay. In sections HQS9-4 and HQS12-5 anisotropic aureoles are weakly developed around quartz grains and a few micritised ooids, and these show strong 1st order yellow, grey and red interference colours depending on the degree of iron saturation. The rest of the weakly developed anisotropic matrix in these slides are predominantly whites, greys and yellows in cross polars, although they become more orange within the organic-rich mottled areas. HQS11-5 also shows 1st order interference colours and has particularly well developed, large iron-rich anisotropic aureoles centred around carbonate grains and some organics.

In all sections, the matrix colour and degree of anisotropism suggests that the 'Btg' of the Grey Clay experienced moderately fluctuating moisture levels. Abundant partially decayed carbonaceous matter and associated limonitic stains have tended to impart some strong colouring in plane-polarised light and again have reduced birefringence under cross-polars. However, the matrix does show a weakly developed microfabric associated with ped formation, suggesting a moderate degree of pedogenesis. The preservation of much organic matter and the dull colour of parts of the matrix under plane polars, suggests that the clays were temporarily or seasonally waterlogged.

5.7.5. The Grey Clay 'BC' horizon

The lowest clay horizon in the soil profile has much less clay matrix and this is a fairly uniform dull khaki-brown to greenish grey in plane polarised light. Abundant khaki-green and dirty brown, heavily micritised grains, which are difficult to differentiate from the clay matrix, are also seen in this horizon (Fig. 5.16e). Larger fragments of altered oolite contain ferrihydrite opaques, which stain the surrounding matrix a deep brown colour (Fig. 5.16e). The clay is also mottled toward the abundant colloidal organic matter, which imparts a brown isotropic matrix (Figs. 5.14e and 5.16e,f).

Much of the clay matrix is almost isotropic in cross polarised light, and although not completely black, is quite dull and heavily stained with colloidal plant matter, carbonate mud and hydrated iron minerals (Fig. 5.14e and 5.16f). The matrix which does show birefringence, exhibits weakly developed anisotropic flecks which are randomly orientated. In some places they have become aligned toward ped surfaces and

coalesce in streaky bright clay argillans. Weak anisotropic aureoles are also seen around some allochems and disseminated quartz grains. 1st order greys and yellows birefringence colours are most commonly observed, although oranges develop around carbonate grains with strongly oxidised weathering rinds.

The very weak anisotropism is completely in accordance with a very poorly developed immature soil horizon overlying a rock regolith. The dull greenish and grey coloration is also suggestive of a permanently water saturated horizon.

5.8. Grain composition and distribution analysis

From the grain size analyses it was established that although the Hornsleasow soils can be described as clay-rich, they are not uniformly fine-grained and contain between 10-60% coarser material depending upon the soil horizon (Figs. 5.10 and 5.11). To investigate the composition and derivation of the coarser material two methods were employed. The first involved sorting sieved fractions, or representative samples of the fractions, into groups depending upon clast types and determining the weight percentage (total rock) of each group. The frequency of each grain type was plotted against grain size to produce a histogram for the four sieved horizons (Fig. 5.17). The second method involved direct point counting and classification of clast types within the soil matrix in a small area at x40 magnification under the petrographic microscope (Appendix C5).

The former analysis recorded all clast types regardless of whether they were organic, inorganic, pedogenic or derived in origin. Clast types recognised in the sieved fractions can be placed into three broad categories. The first are those clasts which are derived from a parent rock. These include rock fragments in the most composed of the oolitic limestone, and detrital or skeleton grains. Most of the skeleton grains are disseminated ooids, bioclasts and quartz ooid-interiors derived from the karstic and pedogenic weathering of the underlying limestones. Fine quartz and alkali feldspar sand grains also occur in the soils and these are considered to have been deposited with the parent clays (section 5.10).

The second category includes clasts which have formed in the soils by pedogenic processes. These include the ferruginous rhizocretions and nodules, phosphatic nodules and possible calcrete nodules (section 5.9). Also included in this category are the clay peds which were sufficiently indurated to remain cohesive during sieving.

The last category includes material which has become incorporated into the clays during pedogenesis by illuviation and other soil processes. Carbonised rootlets and organic matter are the best examples. However, the vertebrate remains and some of the well-preserved biodebris were also introduced into the clay during soil formation. This latter category is not easily defined and is considered subordinate to the other groups.

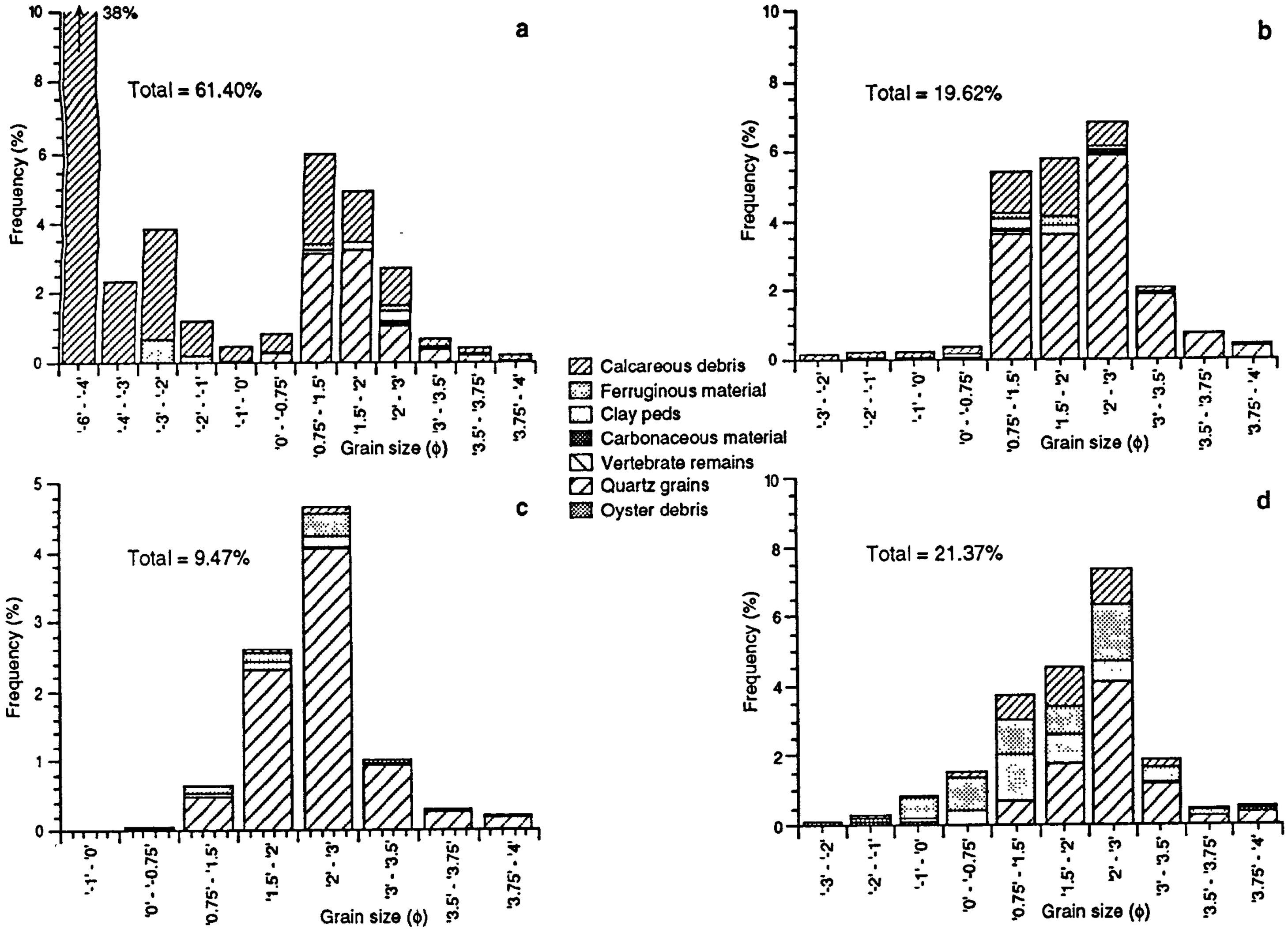


Figure 5.17. Histograms showing composition of individual grain-size fractions for the (a) 'BC' and (b) '2Btg' horizons of the Grey Clay, and (c) 'Btg' and (d) 'Aac' horizons of the Green Clay.

Grain composition and relative proportions in the thin sections was also attempted and again the clasts were split into the ?three categories for easy definition. It was much easier in thin section to establish the amount of non-ooid quartz in the clays, as the these grains displayed shattering associated with a volcanic origin, whilst ooid interiors were composed of detrital quartz. In thin section other features such as grain angularity and preservation could be studied, as well as signs of weathering and diagnostic alteration products. This was particularly important in discerning the relative aggressiveness of the soil pH toward allochems and the preservation of organic matter in the soil horizons. Grain weathering and preservation is discussed in section 5.10.

The distributional texture and degree of sorting was also attempted in the study of the thin sections. Retallack (1990) has suggested four main grain textures which can be applied for soils. These are: (i) *'granular'*, which indicates that the grains are in point-to-point contact, with little or no intervening soil matrix; (ii) *'intertextic'*, where grains are in contact, but where there is occasional intergranular matrix; (iii) *'agglomeroplasmic'*, where grains are fairly rare and the fine-grained material forms localised pockets and an incomplete matrix to the clasts; and (iv) *'porphyroskelic'*, where rare grains are seen to 'float' within the finer-grained matrix.

5.8.1. The Green Clay 'Aac' horizon

Up to 20% of the Green Clay 'Ae' horizon is fine to medium sand (between 1-3ø). The histogram of the coarser fractions shows a fairly well-sorted distribution with a slight coarse skew (Fig. 5.17d). The internal distribution of clast types is mixed containing abundant fine quartz sand (between 10-15%; Appendix C5) and subequal proportions of ferruginous material, crumb-like clay peds and calcareous debris (Fig. 5.17d).

Detrital grains are the most common grain types within the 'Ae' horizon, they consist of the fairly abundant quartz sand and the less common carbonate material. The sorted quartz sand can be split into two groups when studied with a hand lens or under the reflected light microscope. The first appears angular to sub angular with a fresh, vitreous lustre, whilst the second has a frosted translucent appearance and is usually subangular to subrounded. The vitreous quartz makes up about 2-3% of all the sand-sized clasts in the sample (Appendix C5), but is usually finer and is confined to the fine and very fine sand categories (3-4ø). In thin section, the two sorts of quartz can be easily recognised as the former usually displays a fine cracked appearance typical of volcanic origin (section 6.2).

The carbonate grains were subdivided in the sieved fractions into those derived from limestone weathering and those which appeared to have been washed into the soil from nearby coastal marine conditions during high tides and storms. In the case of most of the carbonate grains this proved a futile task as ooids, peloids and bioclastic debris

could have been introduced into the soil at any time. However, the 'Ae' horizon was the only horizon in the soil profile that contained abundant fresh oyster shell fragments. These were clearly not derived from the underlying limestones as they showed no adhering micritic cement and no evidence of karstic weathering. They may have been introduced into the top soil by illuviation of coastal washover, as marine conditions returned (cf. 'oyster hash' in modern coastal marsh soils: Hine *et al.*, 1988). In thin section, shell fragments make 2-3% of the slide, with disseminated ooids and peloid grains being subordinate (1-2%) (Fig. 5.13a,c). Oolite fragments are only present in section HQS10-1 and this sample is rather poorly sorted with an agglomeroplasmic to intertextic distribution of fine to coarse sand (Fig. 5.13a). HQS9-1 has a moderate sorted agglomeroplasmic grain fabric with fine to medium sand grains being represented (Figs. 5.13c and 5.15b). In section HQS12-1 quartz sand is the only detrital grain to be found and the section shows a very well sorted porphyroskelic grain distribution (Fig. 5.13b).

The pedogenic clasts are composed of the ferruginised crumb-like peds and irregular fragments of limonitic material. These are associated with rootlets and organic matter and a few of the larger fragments in hand specimen were identified as parts of fine rhizocretions. Organic matter itself makes up around 1% of the 'Ae' horizon (Fig. 5.17d). Structured cuticle, limonitised rootlets and unstructured carbonaceous inertinite fragments are all found in the thin sections. Charcoal was not found in either the sieved fractions or thin sections. Vertebrate material is extremely rare in all the soil horizons (Fig. 5.17) and in the 'Ae' horizon made up only 0.05% of the material sorted.

5.8.2. The Green Clay 'Btg' horizon

In terms of size distribution of clasts, the 'Btg' horizon was extremely laterally variable over the excavated section. For instance, the sieved sample taken from the edge of site A contains only 10% very fine to medium sand (Fig. 5.17c), but some of the thin sections (notably HQS9-2, HQS9-3 and HQS10-1) contain abundant detrital carbonate sand (up to 20-25%; Appendix C5). Oddly, these samples occur within the centre of the lens (Fig. 5.6), where the clay profile is the thickest (Fig. 4.2), whilst the samples taken from the blocks S11 and S12 located near the sharply rising northern edges of the limestone hollow (Figs. 4.2 & 5.6), have a similarly low proportion of sandy material as of that in the sieved sample. In both the sieved residue and petrographic sections, fine quartz sand (between 1-3.5ø) makes up the most abundant fraction of the sandy grades (between 10-15%; Fig. 5.17c & Appendix C5). Of this up to 5% is shattered volcanic quartz, reaching 10% in section HQS10-3 (Fig. 5.13f). The sieved residue and the sections taken from blocks S11 and S12 show a well sorted normal distribution (Fig. 5.17c) and a porphyroskelic grain fabric in thin section (Figs. 5.13d,e and 5.15d-f), whilst those from the middle blocks are moderately well sorted, with predominantly agglomeroplasmic

distribution (Fig. 5.13f), although this can become intertextic near to degraded oolite rock fragments (Appendix C5).

No carbonate material was found in the S11 and S12 block thin sections, and carbonate detritus is also rare (<1%) in the sieved fraction. However, the middle sections contain up to 10% carbonate detritus. Most of this is fine sand composed of individual bioclasts, peloids and ooids, although some thin sections (HQS9-2) showed up to 5% coarse oolite granules which were badly corroded. Ferruginous and organic matter was rare in the sieved fraction, although sand sized charcoal fragments were discovered, and in thin section limonitised rootlets and organic material were fairly abundant (between 1-5% of the section). Much of this was unstructured carbonaceous material with associated limonitic opaques, although rare cuticle was also recorded in thin sections HQS9-2 and HQS9-3 (Appendix C5). Other material discovered in the sieved residue included rare crumb-like, but non-ferruginised, clay peds and extremely rare bone (0.01%).

5.8.3. The Grey Clay 'Oe' and 'E/B' horizons

No sieved residues were prepared for the 'Oe' horizon of the Grey Clay, therefore all grain composition data is taken directly from thin section analysis (Appendix C5). These show a high proportion of organic matter, up to 10% in section HQS12-4, which is composed mostly of unstructured carbonised plant debris, with limonitic haloes. HQS12-4 also contains structured cuticle and tree resin (Appendix C5). The thin sections have an agglomeroplasmic grain fabric and are moderately sorted, containing fine to medium sand-sized grains of detritus (Figs. 5.14a and 5.16a). This includes abundant quartz sand (10-15%), of which, 2-3% is shattered. The thin section HQS12-4 contains the 'pea-grit' aggregates which have a well-sorted granular grain distribution (Figs. 5.14b and 5.16b) and are almost exclusively composed of fine subangular to subrounded quartz sand (90%), although the occasional bioclast is seen (Appendix C5). The exclusion of the relatively soft and degradable carbonate grains in these aggregates supports differentiation by animal activity.

Carbonate grains are much more abundant in the other sections and can make up to 15% of the rock (HQS10-4); and again there is a distribution difference between the centrally located block S10 and the section taken from S12, where allochems are much rarer (3-5% in section HQS12-4). The abundant carbonate grains in HQS10-4 include coarse sand particles of oolite, as well as disseminated allochems. In contrast HQS10-4 contains only fine to medium sand-sized ooids and some bioclasts (Appendix C5). No other grains were recognised in the sections.

5.8.4. The Grey Clay '2Btg' and 'Btkg' horizons

The Grey Clay 'B' horizon is slightly more sandy than the overlying 'Oe' horizon, although the sieved residue from the original clay lens contains only around 20% fine-medium sand (1-3ø). In the thin sections the distribution of sandy material is quite uniform, with agglomeroplasmic (Fig. 5.16c) and intertextic (Figs. 5.14c,d and 5.16d) matrix supported grain fabrics being noted (Appendix C5). The histogram (Fig. 5.17b) is quite similar in shape to those for the Green Clay, particularly the 'B' horizon (Fig. 5.17c), with a moderate sorted normal distribution, but there is more carbonate material in the Grey Clay (Fig. 5.14d). This material includes up to 5% coarse sand and granules (1 to -3ø) of oolite, and the thin sections range from being poorly sorted at the edge of the excavated site B (HQS11-5 and HQS11-5; Fig. 5.6) to very poorly sorted in the centre (HQ9-4), where granules and small pebbles up to 2-4mm (-1 to -2ø) in width are found. The rest of the carbonate matter is disseminated allochemic material, making up about 5-10% of the rock (Fig. 5.17c & Appendix C5).

The histogram shows an abundance of fine quartz sand (15%) and this is also apparent in the thin sections, were the concentration can reach 20% (e.g. HQS9-4). The volcanic shattered quartz is present in this clay horizon too, and makes up between 3-10% in the thin sections and sieved residue. Well-preserved carbonaceous plant matter is reasonably abundant in the sieved residue (0.5%) and within the sections, and includes medium sand sized grains of charcoal. In thin section the unstructured carbonised plant matter makes between 5-7% of the surface studied and is commonly associated with limonite-stained haloes (Fig. 5.14c). Ferruginous nodules and root casts make up less than 1% of the sieved residue. Structured cuticle and resin fragments were identified in the thin sections and included a large ?megaspore in section HQS9-4. Clay peds and vertebrate remains were very rare in the samples, and make up less than 1% in total of the sieved fraction.

5.8.5. The Grey Clay 'BC' horizon

The basal 'BC' horizon of the Grey Clay is extremely rich in detrital carbonate material and the histogram of the sieved fraction shows a clear bimodal distribution (Fig. 5.18a). This is mirrored in the thin section which is very poorly sorted, with a granular to intertextic grain distribution pattern. The sieved residue contains 61% fine sand and above, but of this, nearly 40% of the weight percentage was made up by a large karstic pebble (-6ø). In fact most of the material above a grain size of 1mm (0ø) is composed of derived oolite from the underlying palaeokarst. Disregarding this coarse fraction, one can see that the distribution of the very fine sand to medium sand fraction (1-4ø) is very similar to those exhibited by both the B horizons of the Grey and Green Clays (Fig.

5.17b,c). Again quartz sand is the most abundant grain composition within this fraction (10-15%; Fig. 5.17a; Appendix C5) and includes around 2-5% shattered grains.

The rest of the sand fraction is composed of a mixture of allochemic grains, with subordinate ferruginous and organic matter. The allochems make up to 15-20% of the rock and include abundant identifiable fragments of oolite, bioclasts, peloids and ooids, in varying degrees of preservation (Figs. 5.14f and 5.16e,f; section 5.10). In the thin section, there is up to 5% unstructured carbonaceous plant matter and this is associated with limonitised haloes. Carbonised debris is also fairly frequently found in the finer fractions of the sieved residues, and includes charcoal pieces. Ferruginous nodules and plant matter were also recovered from the sieved residue and included quite large fragments, up to 8mm (-3ø) in diameter (Fig. 5.17a). Other grains are rare but include fine crumb-like clay peds and about 0.1% bone (which is higher than for the other sampled horizons).

5.9. Pedogenic clasts - glaebules, mottles and nodules

Pedogenic clasts are those naturally segregated lumps of soil matter, which are formed *in situ* within the soil and are commonly called 'glaebules' (Brewer, 1976). Glaebules, in common with soil cutans, are composed of a variety of materials and some examples include calcareous, manganiferous, ferruginous, phosphatic and clay nodules, and faecal pellets. In contrast to cutans, glaebules are not planar and usually have a distinct outline and shape. Glaebules are extremely variable in shape and internal structure though, with nodules ranging from sub-spherical to tubular and highly irregular. They are usually described by their internal composition.

Glaebules can be subdivided into three groups, which are nodules, concretions and mottles. The internal structure of the pedogenic clast is the best criteria for distinguishing between these first two features, as nodules are usually uniform internally, whereas concretions show a concentric structure of 'growth rings' (Retallack, 1990). It is thought that nodules are formed by continuous differentiation and growth of the mineral and these include the calcareous nodules found in desert soils (Retallack, 1985). Septarian nodules also have a system of radiating cracks within the interior, which form when the nodule changes volume. This is very common in wetland soils were sideritic or ferroan carbonate nodules, on exposure to air, dry out irreversibly and septarian cracks form (Retallack, 1990). By contrast concretions are formed by discontinuous, even seasonal growth and common examples include ferruginized concretions found in seasonally wet soils (Retallack & Dilcher, 1981). Both nodules and concretions have a distinct outer boundary with the enclosing soil matrix and are indurated, whereas mottles represent diffuse patches of segregated material or colour discoloration within the matrix. Mottles are composed of the same materials as other glaebules and are considered to be

the earliest stage in glaebule segregation. Good examples of mottles are present within the Hornsleasow soils, for example the limonitic stained yellow mottles in the 'A' and 'Btg' horizons of the Green Clay and the drab mottles of the 'Oe', 'E/B' and 'B' horizons of the Grey Clay.

The composition of a soil glaebule will reflect upon the localised conditions within the profile surrounding the feature, provided there is good evidence that the glaebule is *in situ* and formed during the life of the soil. Glaebules are not formed exclusively in soils, and there are many different types which can be formed in other sedimentary regimes. For instance, manganese nodules can form on the deep ocean floor, and calcareous nodules form in some lacustrine conditions and around hot springs (Retallack, 1990). Therefore, care must be taken to determine whether the feature formed in the soils' parent rock, in the soil itself, during diagenesis and deep burial, or even in an entirely different sedimentary regime being later transported to the soil. The best methods for recognising glaebules which formed *in situ*, is to first establish their relationship with other soil features, say if rootlets avoid an indurated glaebule then it is likely that the feature was present within the soil, and secondly to determine their chemical compatibility with the soil material around them (Retallack, 1990).

In the Hornsleasow paleosols there were at least three different types of possible pedogenic clasts. These are described below and their origins and bearing upon the chemical conditions of the soil discussed. In outlining the palaeoenvironment of the Hornsleasow soils the compositions of the different glaebules are shown on Fig. 5.18, which shows stable mineral associations within common soil conditions of oxidation (Eh/mV) and acidity (pH) (after Retallack, 1990).

5.9.1. Mineralogy and genesis of ferruginous glaebules

Ferruginous nodules and concretions are some of the most common glaebules within the soils, these occur either as ferric iron oxides (haematite Fe_2O_3) or hydroxides (limonite $FeO[OH].nH_20$ and goethite FeO.OH). Sideritic ($FeCO_3$) glaebules have also been identified from the soils (Vaughan, 1989), but this study has failed to discover any, although ferroan carbonate clasts are found within the clays (section 5.9.3). Pyrite is also found in the soils, but not as discrete pedogenic clasts (section 5.9.4). The co-existence of so many different mineral forms and oxidation states of iron (Figs. 5.12 and 5.18), suggests a variable environment for the paleosol units, or one in which there were many microenvironmental niches developed.

The most obvious ferruginous glaebules in the Hornsleasow paleosols are the ferruginous rhizoretions (section 5.4.3), the limonitised yellow mottles and the limonitic-replaced plant matter (section 8.3.3). Ferruginous nodules where commonly recovered from the sieved extractions and some are seen to display fine septarian cracking

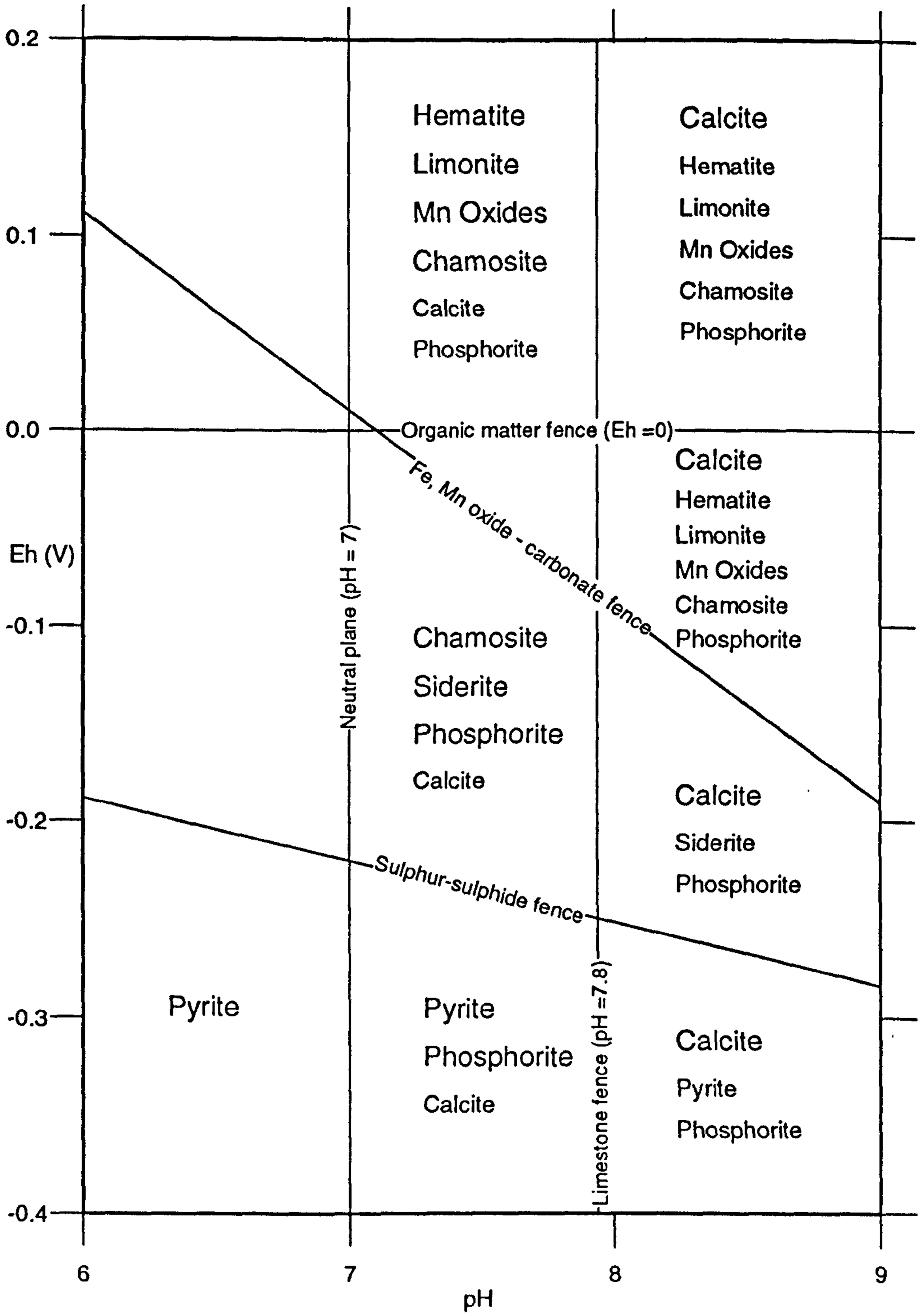


Figure 5.18. Stabilty fields for various mineral associations within the Hornsleasow palcosol profile (modified after Retallack, 1990).

when sectioned. Also commonly observed amongst the sieved fractions are organic material replaced or infilled with ferruginous minerals. Much of the carbonised plant matter is associated with limonitic replacement, but also bones and invertebrate shells may be infilled with limonitic material. The most common example of this are the freshwater gastropod fossils, which may only show a patchy dusting of carbonate ?shell material surrounding a solid ferric oxyhydrite cast (section 8.2.5).

In the first case, dissolved iron that has been mobilised in the drab ferrous state within the reducing seasonally wet soil can be oxidised near to living rootlets to yellow and red ferric oxides which precipitate as concretions around the rootlets (Fig. 5.12; Retallack, 1990). Some of the Hornsleasow rhizocretions contain remnants of an organic cortex within their interior (section 5.4.3). Although it is fairly conclusive that the ferruginised root traces, rhizocretions and associated mottles were formed by processes within the soil environment and that their formation has bearing upon the conditions within the upper Green Clay soil (at least), it is not so easy to prove that the other glaebules were formed *in situ*.

Wherever there is readily available dissolved iron within fresh-water systems,. bacterial action upon decaying plant material produces a coating of iron oxides and oxyhydroxides (Spicer, 1991; section 8.3.3). It is highly probable that this process was responsible for the high proportion of limonite coatings upon the largely undecayed plant matter and vertebrate remains within the waterlogged Grey Clay soil. The same process may have been responsible for ferric oxide/hydroxide replacement and infilling of aragonitic freshwater gastropod shells within the soil (section 8.2.5). It is likely that the organic matter coating and replacements where initially some form of ferrihydrite, but diagenetic alteration of pyrite during burial may also be responsible (section 5.9.4).

5.9.2. Mineralogy and genesis of phosphate nodules

Phosphatic nodules occur within the Grey Clay unit of the Hornsleasow paleosol profile, these are usually irregular indurated objects and in some cases contain moulds of tiny freshwater invertebrates, such as gastropods (section 8.2.5) and/or septarian fractures within them. The phosphatic nodules which occur within the Grey Clay are formed by dissolution of unstable primary biogenic phosphate, hydroxyapatite ($Ca_{10}[PO_4]_6[OH]_2$), in the form of bones, teeth and coprolites, and re-precipitation as a more stable phosphatic compound. Biogenic hydroxyapatites, have a poor crystalline structure which means that they are 10^4 times more soluble than inorganic hydroxyapatites and phosphorites (Fig. 5.18; Lucas & Prevot, 1991). In most terrestrial environments bone material is not known to survive more than 10, 000 years (Newesley, 1989) as it is readily broken down within acidic soils (<pH5, Fig. 5.20; Behrensmeyer, 1978, Retallack, 1984).

Dissolution of the primary hydroxyapatite is brought on by the acids produced by microbial action, and occurs as the organic constituents of bone are also broken down by decay (Lucas & Prevot, 1991). As dissolution of the biogenic hydroxyapatites occurs new phosphatic minerals are precipitated in their place, so that in most cases this fossilisation process does not destroy the texture of the bone material. Depending upon the pH of the surrounding soil, the replacement minerals will be inorganic or secondary hydroxyapatite, at pH 7-7.5; octacalcium phosphate, at pH 6-7; or brushite ($CaHPO_4.2H_2O$) at pH 4.5-6 (Lucas & Prevot, 1991). Where there is excess iron in the sediments and the conditions are fairly acidic, the blue mineral vivianite ($Fe_3[PO_4]_6.8H_2O$) can be formed. However, within freshwater systems such as soils the natural concentrations of the element phosphorous (P) are much lower than in marine water (Prevot & Lucas, 1990) and therefore, apatite (flouroapatite $Ca_{10}[PO_4]_6F_2$) or chloroapatite $Ca_{10}[PO_4]_6Cl_2$) itself is unlikely to form (Lucas & Prevot, 1991).

The XRD analysis of the fossilised bones, teeth, and coprolites and the pedogenetic phosphate nodules at Hornsleasow show that they are composed predominantly of inorganic hydroxyapatite, indicating a pH of between 7-7.5 (Lucas & Prevot, 1991). This is clear evidence that further phosphate dissolution must have taken place within the soils to have produced the inorganic hydroxyapatite nodules. These are seen to nucleate around originally non-phosphatised fossils such as aragonitic gastropod shells (section 5.2.5). Phosphatisation of such fossils has been demonstrated to be caused by bacterial, fungal or algal action (Prevot & Lucas, 1990, Lucas & Prevot, 1991). This rare process occurs in localised acidic microenvironments which where in close proximity to organic phosphate liberated by decay, and the dissolution of the shell carbonate causes apatite to form (Fig. 5.18). Lucas & Prevot (1991) consider six factors which can control the formation of phosphates within a sediment:

(1) There must be abundant dissolved phosphorous within the system
(2) The phosphorous must be able to accumulate within the sediment
(3) The environment must be fairly acidic (or locally acidic) and there must be low calcium carbonate levels (Fig. 5.18)
(4) The conditions should be fairly anoxic (or locally anoxic) (Fig. 5.18)
(5) There must be high levels of fluorine (to form fluoroapatite)
(6) There must be low levels of magnesium within the sediments.

These conditions are satisfied within the Hornsleasow Grey Clay soil microenvironment if the unit is considered to have been formed within a waterlogged conditions. Firstly, both phosphorous and fluorine are abundant within the vertebrate remains. Bones and coprolites are much more susceptible to degradation by micro-organisms than teeth, this

is because the former are much more porous and the latter have a compact outer covering of relatively inert enamel (hydroxyapatite, with little organic matter) (Prevot & Lucas, 1990). Within the sieved fractions some bones, teeth and coprolites show evidence of further dissolution upon their surfaces, which appears to be related to the localised effects of humic and soil acids. The poor preservation of some of the carbonate invertebrate fossils and the good preservation of the wood material within the Grey Clay horizon suggests that in places acidity could have been quite high, although this would have been buffered by the surrounding limestone 'walls' to the hollow. Anoxia associated with organic degradation is also suggested by the presence of pyritised fossils. Magnesium is fairly low within both paleosol horizons (Table 5.5).

Finally, sedimentation rates are considered to be fairly high within the Grey Clay times and there was a constant 'rain' of vertebrate cadavers and isolated remains into the hollow during this period (Metcalf *et al.*, 1992). This rain and the churning effect of the basal mud by animal activity would have buried much material and effectively could have produced a closed microenvironment where microbial decay and acid erosion of phosphatic vertebrate matter could take place. This would produce phosphate-saturated interstitial pore-water resulting in direct precipitation of phosphate and replacement of carbonate fossils.

5.9.3. Carbonate nodules

Both paleosol units at Hornsleasow contain abundant carbonate in the form of skeletal grains and rock fragments. Fairly soft, massive nodules of carbonate (Fig. 5.8) were also discovered in some sections of the lens, and particularly within the Grey Clay. These irregular features appear to be composed of a white soft powdery micritic or microsparite matrix. Their reaction with acid ranges from fairly subdued to rather vigorous and some are finely dusted with a limonitic coating. It is possible that both ferroan and non-ferroan carbonate nodules are present within the soils. Some exhibit fine septarian cracking and show displacive fabrics within the soil matrix suggesting that they formed *in situ* within a seasonally waterlogged soil, but none seem to be related to rootlets as carbonate rhizocretions. In thin section, the nodules show abundant quartz silt and organic inclusions, which suggests that they formed as displacive carbonate glaebules (Wright & Peeters, 1989) rather than as sparmicritised clasts of oolitic debris (Brewer, 1976). Most of this fine micritic carbonate occurs within the 'Btkg' and 'BC' horizons of the Grey Clay.

Calcareous nodules form in two environments at the present, the first are the dry desert soils where calcareous material is leached from the surface but there is insufficient rainfall to deplete subsurface concentrations. These soils are characterised by indurated layers of subsurface carbonate called 'calcrete' (Allen, 1986) and show abundant

rhizocretions. The second environment occurs in wetter regions, which are coastal or karstic regions, where soils contain an over-abundance of calcareous material (Retallack, 1990, Davies, 1991). In these soils repeated wetting and drying will cause alternating localised acidic and alkaline conditions, which will leach, mobilise and re-precipitate the carbonate in turn. This would appear to be the case in the Hornsleasow paleosols and in particular the wet Grey Clay. This is supported by the strong evidence of carbonate dissolution and micritisation of skeletal fragments within both clay units (section 5.10). That seasonal desiccation continued to effect the nodules after formation is illustrated by brecciation of the clasts, in the form of synsedimentary septarian cracks and gashes.

The formation of the micritic, siderite and ferroan calcite glaebules is related to dissolution and replacement of biogenic carbonate skeletons within the clays and dissolution of calcitic cements and bioclasts of the surrounding karstified limestones of the Chipping Norton Formation within the fresh-water system. Large quantities of organic matter, such as plant and animal material, are degraded by 'methanogenesis', rather than by sulphate-reduction, particularly within anaerobic conditions (Canfield & Raiswell, 1991a), which proceeds as:

$$\text{Organic matter} + \text{bacterial degradation} = CO_2 + CH_4 + NH_3 + 2H_3PO_4 \qquad 5.8$$

This reaction and root respiration within the surface horizons of the soil produce carbon dioxide (CO_2), which can acidify the pore-water, so that the sediments become undersaturated with respect to calcium carbonate. Localised build-ups of soil CO_2 would have occurred in the Hornsleasow clays, because of the relative impermeability of the clay matrix, which would also have provided a limiting factor upon wholesale dissolution of the underlying limestone, as free porewater circulation from the CO_2 producers to the bedrock was reduced.

Organic acids and chelating products such as humic acid, are produced during both growth and decay of plant material and these further serve to acidify the soil pore-water, particularly at the base of the waterlogged and anoxic Grey Clay. The acidic conditions would form a contributing factor in enlarging the solutional hollows in the limestone (J. Cole, pers. comm). This acidity would account for the excellent preservation for plant fossils in the Grey Clay (section 8.3.3) and the fact that some of the microvertebrate remains and coprolites exhibit localised dissolution upon their surfaces (Metcalf & Walker, 1994; section 10.6).

5.9.4. Pyrite - pedogenic or early diagenetic?

The iron sulphide, pyrite (FeS_2), is rare in both clay units, although disseminated tiny framboids, clusters and aggregates of pyrite are found within, associated with and

occasionally replacing organic remains (Chapter 8). However, no isolated, non-organic glaebules of the mineral have been found within the sediments.

The formation of pyrite is understood to take place in sediments below the zone of oxic respiration and decomposition over a range of natural pH (Figs. 5.12 & 5.18), as the result the activity of sulphate-reducing bacteria during the first stages of decay of organic matter (Berner, 1970, 1984). The process of formation is known to take place over a number of stages but generally occurs via the reduction of sulphate (SO^{4-}) to produce hydrogen sulphide (H_2S) (Allison, 1990) and the reaction of sedimentary iron (Fe^{3+}) in the form of iron oxides and hydroxides with the hydrogen sulphide to produce iron sulphides (Canfield & Raiswell, 1991b). These reactions can be summarised neatly by the following equation (after Berner, 1970, 1984):

$$18CH_2O + 9SO_4^{2-} + 6FeOOH = 18HCO^{3-} + 3FeS + 3FeS_2 + 12H_2O \qquad 5.9$$

The favourable geochemical conditions required for pyrite growth can be summarised by a flow chart (Fig. 5.19). Hydrogen sulphide production is limited by the nature of the organic material within the sediments and to a lesser extent, by the concentration of dissolved sulphate, whilst the amount of pyrite production itself is limited by the availability and mineralogy of the detrital iron phases within the clays (Canfield & Raiswell, 1991b). In non-marine systems the paucity of sulphate within the pore-water is the limiting factor on the formation of pyrite even within iron-rich, rather anoxic sediments and decay of organics tends to be controlled by methanogenesis rather than sulphate reduction (Canfield & Raiswell, 1991b). Nevertheless disseminated pedogenic iron-bearing minerals are abundant within the limestones and clays at Hornsleasow, and dissolved iron in its drab reduced-state (Fe^{2+}) would accumulate (Canfield & Raiswell, 1991b). The source of the ferrous iron is probably by reduction of sedimentary iron oxides by organic compounds, probably mediated by bacterial decay processes (Canfield & Raiswell, 1991b) in acidic environments (Fig. 5.12), for example the reaction:

$$CH_2O + 4FeO[OH] + 8H^+ = CO_2 + 4Fe^{2+} + 7H_2O \qquad 5.10$$

Canfield & Raiswell (1991b, p.352) also suggest several other bacterial processes involving sulphate-reduction which could produce abundant dissolved iron. Ferrous iron is extremely reactive and will readily combine with the anions such sulphide, bicarbonate (HCO^{3-}) and phosphate which are the metabolic products of microbial decomposition to produce precipitates of pyrite, siderite and vivianite respectively (Fig. 5.12; Canfield & Raiswell, 1991b). The conditions required to generate pyrite growth within these "iron-dominated" (Canfield & Raiswell, 1991b, p.380) pore-waters can be extremely localised

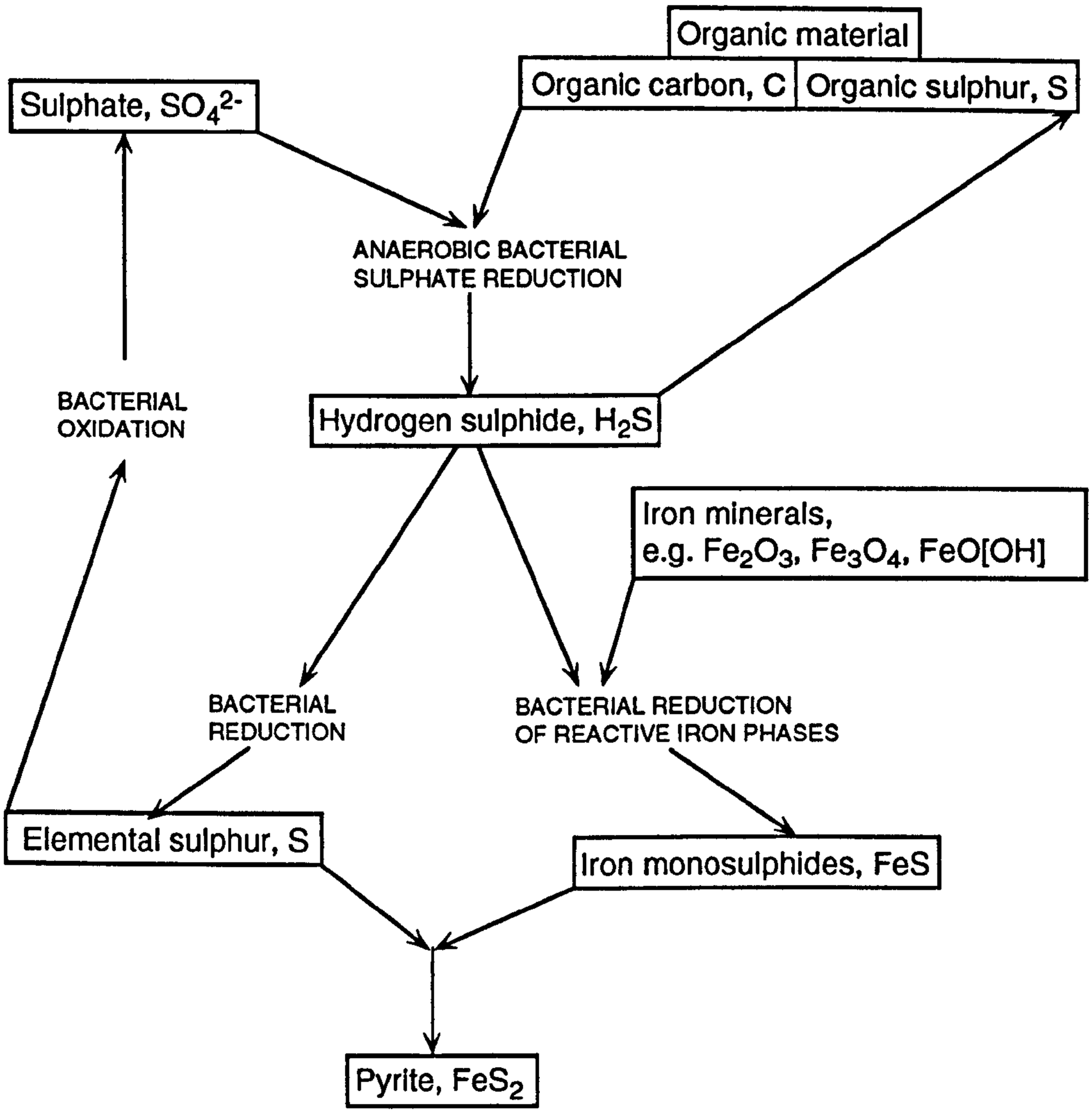


Figure 5.19. Schematic representation of the geochemical pathways and processes involved in the formation of sedimentary pyrite by sulphate reduction (after Berner, 1972, Allison, 1990, Canfield & Raiswell, 1991a).

and even at the scale of cracks and surface depressions within organic material (Chapter 8). Favourable geochemical conditions are usually concentrated at the site of decomposing organic carbon generating hydrogen sulphide which could fix upon the pore-water supply of iron (Berner, 1984). At Hornsleasow, most of the pyrite concentrations are found within the interstitial parts of bone and wood, and also occur as sedimentary infills within small calcitic gastropod shells. However, the complete replacement of organic matter by pyrite is extremely rare, as only one small rootlet within the 'Bt' horizon of the Grey Clay (Fig. 5.11j,k) was found to have been permineralised by early diagenetic pyrite. That more material was not pyritised suggests unfavourable geochemical conditions or diagenetic alteration.

Diagenetic alteration of the pyrite to iron oxyhydrates cannot be ruled out entirely, but a more plausible explanation is that the limonite was precipitated as a disseminated decay product with the plant matter (section 5.9.1). Limonite is the principal iron mineral in freshwater terrestrial systems, forming especially in the organic-rich, waterlogged soils with fluctuating oxygen levels (FitzPatrick, 1993).

The presence of early diagenetic pyritous replacements of some fossils might suggest surface water gley within the lower Grey Clay paleosol and this is supported by the preservation of wood as carbonised traces with dark grey 'drab-haloes' (Fig. 5.7). Drab-haloes are a type of glaebule which are often seen around rootlets and organic matter in paleosols (Retallack, 1990). Drab-mottles are associated with surface reduction and gleying, and are often found in lowland regions with impermeable clay-rich subsurface horizons. Pyrite is a common mineral constituent of such glaebules (Retallack, 1990), and this could be readily oxidised to ferric oxyhydrites and oxides upon burial (Walker, 1967). This type of drab mottling commonly occurs within clay-rich, seasonally waterlogged soils and is associated with the reduction and mobilisation of iron by anaerobic bacterial activity in the stagnant water surrounding the rhizosphere or root zone after decomposition of the organic material. Complete pyritisation of the rhizosphere can occur, typically in soils formed within fairly reducing, acidic and brackish-waterlogged conditions occuring in estuarine environments (Fig. 5.18; Retallack, 1990). The sections taken from the 'Oe' horizon of the Grey Clay, contained abundant opaque material, but this proved to be a mixture of carbonaceous matter and yellow limonite stains, rather than pyritic (section 5.7.3 & Appendix C5).

Drab-haloes within organic-rich horizons can also form from organic matter 'staining' of the soil during burial, irrespective of reduction state of iron and waterlogging of the horizon (Retallack, 1990). This staining is caused by the anaerobic bacterial breakdown of buried woody tissue, to form dispersed organic matter or 'kerogen'. Organic-induced drab-haloes can be identified by checking for comparable amounts of total iron (Fe^{3+} and Fe^{2+}) within the drab and normal sediments within the

soil horizon and this would seem to be the case within the Hornsleasow Grey Clay paleosol surface horizon (section 5.6).

5.10. Weathering profile throughout the clays and soil forming processes

Soils are in essence the accumulated stable weathering products of rocks and minerals under normal atmospheric conditions and usually form over a period of between 1000 and 1 000 000 years (Retallack, 1990). Hence, the processes involved in their formation and maturation are the different types of weathering acting upon the parent material. Soils are formed by the interaction of the three types of surface weathering: physical, chemical and biological. These distinctions are fairly arbitrary as a process may proceed by a combination of these types, for instance roots leach the soil of its nutrients (biological weathering), but they also exert a physical pressure on the confining soil matrix and they change soil conditions around them by chemical reactions, such as the production of organic acids.

The recognition of different soil-forming processes in a paleosol profile, can help elucidate their meaning and relationships, providing an invaluable insight into the palaeoenvironment in which the soil matured. In soil science, the procedures for studying soil weathering include determining the pH of soil solutions, soil organic matter concentrations and the linear expandability of the clay minerals within the soil matrix by direct measurement (Retallack, 1990). None of these parameters are directly measurable in paleosols which may have been altered by burial and diagenesis. The three weathering types are defined below and features identifiable in paleosol investigations described for the Hornsleasow fossil soils.

5.10.1. Physical weathering

The physical destruction of a parental rock or sediment to form a soil, can occur by a variety of processes, such as freeze/thaw relationships, wetting and drying, and fire heating. Most of these processes leave characteristic features within the soil structure. Soil 'loosening' is the general term for physical weathering processes which cause rocks or soil to fragment. Jointing in rocks and the open pore-space between peds in soils are good examples of physical loosening, both of which can be affected by compaction and cementation after burial. Joints and porosity are the major routeways exploited by solutional and other erosional processes (Trudgill, 1985). The ped structure can give an approximation on the original porosity of the soil, and the type of fluid flow. Permeability can be estimated from the main mineralogy and grainsize of the soil matrix.

The Hornsleasow paleosol units are both clay-rich and upon excavation were found to retain much moisture. Although this fluid is unlikely to be the original soil moisture, the retention of any fluid at all, suggests that these clays have a high 'field

capacity' for fluid-flow. Field capacity is the amount of soil fluid which can be retained in micropores, once the large interped pores have been squashed or infilled during diagenesis (Retallack, 1990). The network of angular blocky peds within the main clay horizons suggests that fluid-flow was quite high in the soils, and particularly within the 'A' horizon, which is rooted and contains crumb-ped development. The presence of ferruginous concretions within the soil profiles, also indicate free-flowing well-oxygenated water. Although, if sideritic nodules are present in the lower Grey Clay, these would suggest stagnation of poorly oxygenated water within a soil of low permeability (Figs. 5.12 & 5.18).

The soil structure and microfabric can give some indication of previous physical soil movement, especially in clay-rich paleosols. Clays, such as smectites, swell and shrink during periodic wetting and drying, giving characteristic domed and columnar ped structures. These were not observed at Hornsleasow, and x-ray diffraction analysis of the clay mineralogy, would seem to indicate a low trace of smectitic layering (see below). The microfabric of a clay matrix can be directed related to the swelling capacity of the dominant clay mineralogy, with highly swelling matrices exhibiting woven or reticulate anisotropic zones, whilst flecked or isotropic matrices suggest a low swelling capacity (Holzhey *et al.*, 1974). In cross-polarised light, the anisotropic microfabric of the clays was dominantly a flecked bright clay, with occasional continuous or fibrous zones, suggesting only a low amount of movement. Slickenslides also suggest clay matrix movement, and are present in the Hornsleasow sections.

Although, freeze-thaw and fire damage cracking of soils, cannot be directly measured, they do leave characteristic features. The presence of charcoal within the fossil soils indicates that forest fires where prevalent in the Bathonian landscape (section 8.3.4), and fire heating may have caused physical movement in the soils.

5.10.2. Biological weathering

Biological modification of soils consists mainly of microbial, invertebrate and plant activity and can effectively regulate the physical and chemical properties of the soil. The three most important biological processes which take place within a soil, are humification, bioturbation and soil nutrient uptake. All of these processes reflect and affect the overall productivity of the soil and can be usually assessed even in diagenetically altered paleosol horizons (Retallack, 1990).

(a) *Humification.* This is the process of organic matter decay and proceeds through a series of biologically controlled chemical reactions, were complex organic molecules are oxidised and broken down into their simpler constituent molecules, (e.g. CO_2), which can be used in the metabolism of other organisms. In physical terms it is the fragmentation of large organic material into an amorphous mass of fine organic

matter. Humification is caused by a wide variety of soil invertebrates, vertebrates, plants and microbes.

The degree of humification can be assessed for paleosols in terms of the amount of identifiable fragments of plant material and indeterminable organic matter. The total organic content (T.O.C.) of the horizons can also be assessed, but this may be modified as volatile organic matter is lost through diagenetic and metamorphic alteration. Erosion of the humic horizon at the surface prior to burial and microbial activity in buried soil profiles can also contribute to the loss of volatiles in paleosols (Retallack, 1990).

In modern day and fossilised humic horizons, there are three types of humus distinguished upon their relative amounts of identifiable to unrecognisable plant fragments. The types are 'moder' humus which is essentially fossil leaf litter, and contains plant remains in varying stages of preservation; 'mor' humus, which shows little evidence of humification and contains a high proportion of intact plant organs; and 'mull' humus, which contains indistinguishable fine-grained colloidal organic matter (Retallack, 1976, Wright, 1983). From point counting organic matter within the sections, it was possible to elucidate the amount of humification which had taken place in the 'Aac' horizon of the Green Clay and the 'Oe' horizon of the Grey Clay. The T.O.C. of the two horizons is less than 5% of the rock for the 'Aac' horizon and around 7-12% for the 'Oe' horizon of the Grey Clay. Almost all of the material in the 'A' horizon has been oxidised to a limonitic stain, or is present as poorly recognisable unstructured inertinite or coalified wood. The 'Oe' horizon of the Grey Clay contains up to 3-4% partially intact plant matter, in the form of fossilised tree resin (resinite), cuticle (cutinite), recognisable spores (sporinite) and charcoal (fusinite) and can be considered to be a fossilised 'humic peat' horizon. The 'B' and 'B/C' horizons of the Grey Clay also contain up to 5% T.O.C. levels, some of which is extremely well preserved (section 8.3.3.)

(b) *Bioturbation.* As the degree of reworking by soil organisms increases within a soil profile, so the physical properties of the soil are modified. For example, constant burrowing and ingestion of soil matrix breaks down larger peds into small crumb-like pellets. This is also facilitated by root activity. Organic sheaths are also produced by soil organisms during this process, which form ped coatings called organans. These sheaths may help to form a cohesive bond between the fine network of peds in well-bioturbated topsoil (Retallack, 1990) and are known to help reduce erosion.

The amount of bioturbation can be quantified by counting the number of roots and burrows per centimetre over a metre squared quadrant (section 5.3). The general features of the soil structure, such as the presence of a crumb-ped arrangement and a bright clay microfabric, are also important indicators of a worked soil.

(c) *Nutrient availability.* Soil organisms are dependant upon a limited number of elements for metabolism. The availability of these nutrients within a soil are

crucial to the productivity of the soil. Plants depend upon the elements, hydrogen (H), carbon (C), nitrogen (N), oxygen (O), magnesium (Mg), phosphorous (P), sulphur (S), potassium (K), calcium (Ca) as macronutrients (Retallack, 1990). With the elements boron (B), chlorine (Cl), vanadium (V), manganese (Mn), iron (Fe), copper (Cu), zinc (Zn) and molybdomen (Mo) being required as micronutrients. They derive these nutrients from atmospheric gaseous exchange and intake of fluids by their roots. Nitrogen, phosphorous and sulphur are taken into the roots as their soluble cationic forms nitrate (NO_3^{2-}), sulphate (SO_4^{2-}) and phosphate (PO_4^{2-}), derived from their microbial oxidation, whilst the other nutrients are taken directly from the soil matrix, by the hydrolytic weathering action of roots (Retallack, 1990). Soil animals also require various macro- and micro-nutrients, which include those utilised by plants and an additional two macronutrients (sodium (Na) and chlorine (Cl)) and nine micronutrients (fluorine (F), silica (Si), chromium (Cr), nickel (Ni), cobalt (Co), arsenic (As), selenium (Se), tin (Sn) and iodine (I) (Retallack, 1990).

Most nutrients which are utilised appear as cationic bases (i.e. positively charged ions) in solution in the soil, therefore in soil science the productivity of a soil is assessed by the amounts of available cations left within the soil fluids. This is not readily assessed in fossil soils although the weight percentages of the bases (such as Ca^{2+}, Mg^{2+}, K^{+}, Na^{+}) within a paleosol can be measured and compared to non-basic cation abundance (such as Al^{3+}) as molecular weathering ratios (section 5.10.3) in an attempt to quantify 'base saturation' in the soil. The base saturation is linked to the acidity of the soil and is further discussed in the next section on chemical weathering techniques. The availability of other nutrients are difficult to assess as they occur in trace abundances within the soil matrix and are mainly recycled by organisms during humification. However, a rough guide to the amount of one of the important nutrients, phosphorous, is the abundance of apatite and bone material within the paleosol (ca. 0.1% soil matrix).

5.10.3. Chemical weathering

Chemical weathering is probably the most easily diagnosed soil-forming process in paleosol profiles. It is largely associated with biological weathering in natural soil conditions and may also give an indication of the amount of physical weathering involved in soil formation and maturation. This study focused upon the recognition of three major elemental reactions and their associated soil forming processes and conditions, which are (a) hydrolysis, dissolution and soil pH; (b) oxidation, reduction and soil Eh; and (c) hydration, dehydration and soil moisture content (Retallack, 1990).

In order to study the effects of chemical weathering, three methods were employed. In the first place, a mineralogical profile of the soils was constructed. This was diagnosed by M.J. Simms (Simms & Metcalf, in prep. 1994) by x-ray diffraction

(XRD) techniques carried out upon glycolated and non-glycolated samples taken from the field (M.J. Simms, pers. comm. 1992). Preliminary studies were also carried out on three randomly taken samples from the Green and Grey Clay in 1988, by a student, Ms. R. Jones, at Bristol University, under the supervision of Dr. D. Robinson. Her results appeared in R.F. Vaughan's account of the clays (Vaughan, 1989), but have not been substantiated by our investigation. The original manuscript, a BSc thesis has also been lost and there was some suggestion that the student may have confused the results or mixed up the samples (D. Dartnall, pers. comm., 1992). Therefore, although her findings and those of M.J. Simms are reported here, it is proposed that in the future a more rigorous profile of the clay mineralogy will be constructed.

The second method involved a careful appraisal of the preservation of skeleton and pedogenic clasts within hand specimen and petrographic thin sections. Quartz grains in particular yield the most information upon chemical weathering within the soils, as the preservation of these will give the top value for degree of alteration. Also important are the amount of dissolution and development of weathering rinds upon limestone clasts and individual allochems in the soil, although much of this alteration could also have occurred prior to their inclusion within the clay soils, during karstic weathering and *terra-rossa* soil formation (section 4.5). Their are five classes of alteration for grain alteration (FitzPatrick, 1984; Appendix C5) are:

Class 0 : Fresh; less than 2.5% mineral altered

Class 1 : Slightly altered; between 2.5-25% of mineral altered and dissolved

Class 2 : Moderately altered: 25-75% of mineral altered and dissolved

Class 3 : Strongly altered; 75-97.5% of mineral altered and dissolved

Class 4 : Completely altered; over 97.5% of mineral dissolved away.

The final method for assessing the degree of chemical weathering, involved a detailed analysis of the major elements within bulk samples of the clays. The bulk chemistry of 35 paleosol clay and mixed clay-sand samples taken from the soil profiles (Fig. 5.8 and C2) was ascertained using atomic adsorption spectrophotometry (AAS) techniques. AAS methods are widely used in the elemental analyses of rock and groundwater (Fairchild *et al.*, 1990), and Appendices D1-3 provide an outline of the experimental procedures carried out upon the 35 samples. The AAS analyses elemental abundance based upon the weight percentage of the oxide of the particular element. Those elements studied and the procedures used to determine their relative abundances in the paleosols are listed in Appendix D4, the results of these analyses are shown in Table 5.5 and Fig. 5.20.

Sample	Weight Percentages							
	SiO_2*	Al_2O_3**	FeO/ Fe_2O_3	MgO	MnO	CaO	Na_2O	K_2O
S9 A1	63.186	18.560	4.996	1.648	0.010	6.320	0.080	3.200
S9 A2	68.529	13.450	3.648	1.208	0.010	10.805	0.070	2.280
S9 A3	62.293	15.050	3.711	1.148	0.013	13.958	0.060	1.767
S9 A4	69.631	8.600	2.491	0.568	0.010	15.383	0.050	1.267
S9 'O'	69.734	14.570	4.591	1.090	0.010	7.325	0.060	1.620
S9 A5	59.099	9.120	2.765	0.842	0.020	24.807	0.060	1.287
S9 A6	64.993	9.120	4.073	0.862	0.050	17.535	0.050	1.317
S9 'C'	18.560	4.570	5.108	0.422	0.050	68.907	0.040	0.343
S10 A1	64.741	19.460	4.994	1.342	0.010	4.223	0.080	3.150
S10 A2	63.455	19.530	5.794	1.978	0.010	3.633	0.090	3.510
S10 A3	66.022	20.330	5.378	1.718	0.010	1.087	0.080	3.375
S10 A3'	64.935	18.770	5.348	1.815	0.010	3.618	0.090	3.414
S10 A4	65.394	12.550	2.791	0.975	0.010	14.700	0.060	1.520
S10 A4'	66.784	12.340	2.931	1.085	0.020	13.250	0.060	1.530
S10 A5	64.264	11.610	3.698	1.105	0.020	15.623	0.070	1.610
S10 A6	49.345	8.110	2.438	0.755	0.060	36.035	0.050	1.207
S10 'C'	22.437	7.660	2.975	0.535	0.070	63.543	0.040	0.740
S11 A1	63.908	20.170	7.184	1.735	0.020	1.467	0.090	3.426
S11 A2	64.612	20.020	5.648	1.595	0.010	2.795	0.080	3.240
S11 A3	63.263	16.510	5.548	1.582	0.010	7.887	0.080	3.120
S11 A4	64.861	10.970	5.481	0.968	0.010	14.150	0.060	1.500
S11 A5	66.645	11.360	2.963	0.972	0.020	14.590	0.060	1.390
S11 A6	52.159	11.830	3.110	0.828	0.030	28.783	0.050	1.210
S11 'C'	28.862	5.170	7.958	0.532	0.040	54.768	0.050	0.620
S12 A1	63.350	19.880	7.652	1.675	0.010	2.093	0.080	3.260
S12 A2	68.605	18.150	5.408	1.552	0.010	0.975	0.080	3.220
S12 A3	68.122	17.870	5.428	1.762	0.020	1.498	0.080	3.220
S12 A3'	69.151	17.460	5.334	1.552	0.010	1.353	0.080	3.060
S12 A4	68.585	17.100	5.222	1.468	0.010	2.647	0.070	2.900
S12 A4'	69.452	17.420	5.098	1.295	0.010	2.338	0.070	2.317
S12 A5	69.412	15.320	4.998	1.115	0.010	5.525	0.060	1.560
S12 A6	67.135	11.360	5.668	0.935	0.012	11.460	0.060	1.370
S12 'C'	23.417	1.030	2.698	0.085	0.020	70.730	0.020	0.000

Table 5.5. Major elemental abundances for the four soil profiles.

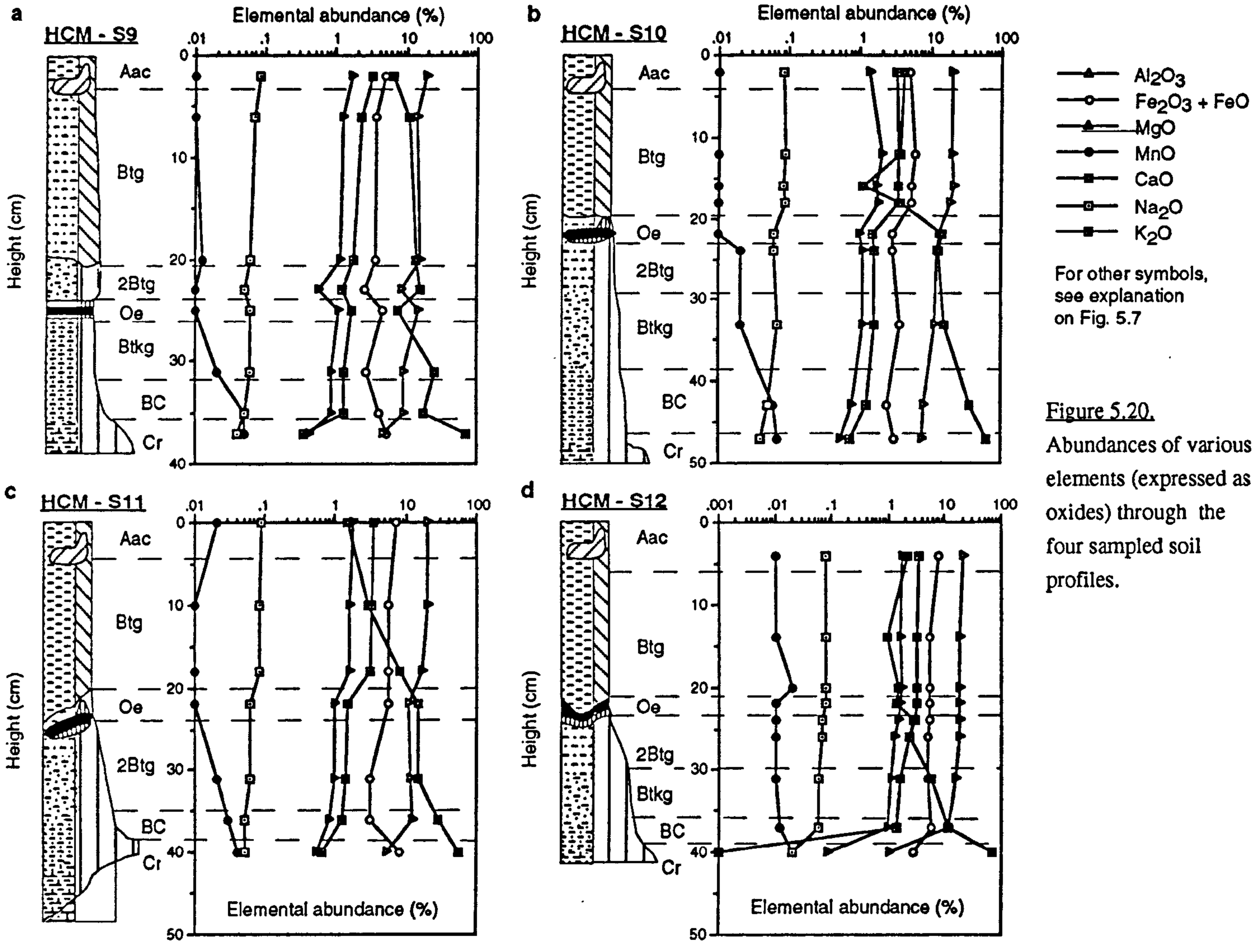


Figure 5.20.
Abundances of various elements (expressed as oxides) through the four sampled soil profiles.

From these results, molecular weathering ratios were calculated to elucidate particular chemical processes within the soil process. Molecular weathering ratios are calculated by dividing the weight percentage of each relevant oxide by its molecular weight and then adding together or dividing as specified by the particular ratio, for the different types of weathering process (Retallack, 1990, table 4.4). There are specific molecular weathering ratios for determining the pH, Eh, hydration, salinisation and leaching of a paleosol horizon (Table 5.6 & Figs. 5.21, 5.22, 5.23).

A more rigorous method of assessing chemical change within a soil profile is to normalise the chemical compositions to a constituent mineral which has remained stable through the transformation of the parent rock into the soil. For this type of analysis one needs to have a fair idea of the potential parent material, so that various assumptions can be made in calculating the weathering formulae (Retallack, 1990, table 4.5). For the Hornsleasow fossil soil profile this was not possible for several related reasons. The first problem is actually finding a suitable parent sediment for the soils, the underlying carbonates are clearly not the prime candidate, and there is evidence that the clays may have been derived from two possible sources, a weathered volcanic ash deposit and/or a terrigenous clay (section 6.1). The only way to elucidate the derivation would be to complete a much more detailed XRD analysis of the clay mineralogy, but even this might fail if the clays had suffered prolonged mixing and weathering within the soil profile or metamorphic alteration during burial (section 5.11).

The results of these studies are outlined below in reference to the three main types of chemical modification of soil material:

(a) *Hydrolysis, dissolution and soil pH.* Hydrolysis and dissolution reactions take place within acidic environments in the soil horizons. Hydrolysis is usually a reaction involving an acid radical (such as carbonic acid, H_2CO_3) with a cation-rich mineral grain, such as a silicate. The reaction produces clay and soluble cations, which may then be leached from the soil by illuviation or plant nutrient uptake (Retallack, 1990). Hydrolysis is the main chemical reaction by which silicate minerals are weathered in a soil (Chesworth, 1973). Dissolution is similar to hydrolysis but involves a mineral disassociating into its constituent ions, rather than producing a new product. The most common dissolution reaction in soils is that of carbonate minerals within a slightly acidic solution (equation 4.1).

In order to ascertain the prevalence of hydrolysis and dissolution reactions within modern soil science, the pH of the soil fluids are directly measured. Obviously, this method is not applicable to fossil soil horizons. Fortunately the original pH of the soil can be estimated from the stability fields of the mineral associations (Fig. 5.18) and their preservation, from mineralogy and molecular weathering ratios. Biological features can also provide indirect evidence for the general acidification of a soil horizon.

The clay mineralogy of the sampled horizons indicates that the Grey Clay was largely composed of a mixed illite and kaolinite fraction (Metcalf *et al.*, 1992), although the earlier analysis has detected the swelling clay smectite (montmorillinite) in association with illite, chlorite and small traces of kaolinite (R. Jones in Vaughan, 1989). The presence of kaolinite, a deeply weathered clay, would tend to indicate a fairly acidic environment with pH as low as 3-4 (Retallack, 1990). A pH as low as this is not indicated by the presence of the fairly easily hydrolysed base-rich illite clay. Illitic and smectitic clays tend to occur in near neutral or slightly alkaline environments together (between pH 6-8). Smectite is a common component of hydrolysed soils where weak drainage, high pH and high evaporation occurs and smectitic clays are particularly base-rich and can occur in extremely alkaline soils (up to pH 9-10). Within stronger acidic environments (<pH 6), smectites will lose Na^+ and K^+ in a process known as 'illitisation' (Retallack, 1990). Complete post-depositional alteration of smectite to illite and chlorite is a common occurrence in Jurassic volcanic ash deposits (Bradshaw, 1975).

The Green Clay, is similarly composed of predominantly illite, with substantially reduced amounts of kaolinite clay (Metcalf *et al.*, 1992). Again the earlier analysis suggested a slightly different composition of illite/chlorite layered clays (Jones in Vaughan, 1989). Jones (in Vaughan, 1989) suggested that smectite may have been present once in the upper clay, but was leached into the lower Grey Clay. This cannot be substantiated until further XRD analyses are carried out on the clays.

The preservation and inclusion of carbonate skeleton and pedogenic clasts within the soils at Hornsleasow is highly suggestive of near neutral to slightly alkaline conditions (Figs. 5.18) and the presence of large amounts of limestone around the clay soils would have provided an effective buffer to acidic conditions. However, the general preservation of the carbonate skeleton clasts within the soils at Hornsleasow is similar throughout the profiles, and suggests a high degree of chemical alteration (Appendix C5). Individual allochems show a variable amount of micritisation along the edges and in many cases have developed a well-formed weathering rind (section 4.4). In some cases, dissolution and decalcification of the carbonate has also occurred, and many allochems contain inclusions of oxic weathering products such as limonite and goethite granules (Appendix C5). Ooids and peloids are usually in the worse condition, with lines of opaque inclusions being present in the inner lamellae of ooid grains and strong opaque coatings. Originally low magnesium calcite oysters and echinoderm fragments are the only recognisable bioclasts within the soils and these usually show dirty brown, dark grey or khaki micritic weathering rinds and opaque coatings (Figs. 5.13-5.16).

In the more carbonate-rich horizons, such as the basal mixed 'BC' and 'Btkg' horizons of the Grey Clay, oolitic clasts are also common. These have also been subjected to significant dissolution, showing decalcification of the sparite and other

cements, with associated grain loss and the development of thick opaque-filled weathering rinds (Fig. 5.14f). It is likely that much of this weathering occurred during the formation of the underlying karstic terrain and *terra-rossa* (section 4.4). These clasts then became incorporated into the overlying clay-rich soils by pedogenic mixing. However, isolated carbonate clasts may have also suffered further chemical erosion within the soils.

Hydrolysis reactions upon silicate grains are best assessed by looking for minerals which are prone or resistant to acidic attack. For instance, quartz is known to be many times more resistant to hydrolysis than feldspar or mafic mineral grains. Neither of the soil units contained any mafic minerals and less than 1% of the felsic grains were alkali feldspar. Therefore this would suggest that these minerals had been leached from the clays at some point. The quartz grains themselves showed variable amounts of dissolution depending upon their origin. Ooid and terrigenous quartz grains were often pitted and showed minor dissolution at high magnification along cracks (Appendix C5). Finer grains were more likely to show weakly developed weathering rinds and the beginnings of stage 1 chemical weathering. On the other hand, the volcanic shattered quartz was quite clear of any signs of hydrolysis (Stage 0; Appendix C5). However, large scale dissolution of quartz grains was not observed, suggesting that they were relativeky stable in the palcosol environment (cf. Foos, 1991).

The only other clasts to show signs of hydrolytic etching are apatite clasts and bone material (section 10.6). Much of the erosion upon the vertebrate remains is likely to have occurred during ingestion, by the action of stomach acids on the bone apatite. However, some examples show evidence of localised acid attack similar to etching resulting from root activity, and soil or organic acids must also be responsible for surface etching seen upon coprolitic and pedogenic clasts. In general though the preservation of much bone material within the soil units is highly persuasive of uniform non-acidic environmental conditions (Fig. 5.18).

It has been previously stated that good preservation potential for wood and plant matter within paleosols is largely a function of high acidic levels (eg Metcalf *et al.*, 1992). Clearly, in modern and Quaternary circumstances the best preserved organic matter is derived from fibric peats, where more than a third of the plant remains are structured and identifiable (Retallack, 1990). However, although many of these peats are acidic, where conditions are hostile to the decomposing microbes, some also occur in Carr or Fen-type marshes, which develop upon carbonate hinterland (Retallack, 1992) and the main preservative measure in these bogs is the exclusion of oxygen and oxic decay by waterlogging (Fig. 5.18). Clearly, the decay of organic matter within the soils would have raised localised acidity levels, but a truly acidic soil is antiseptic and quite often devoid of normal organic activity. The pervasive rootlets in the Green Clay and the

bioturbation at the surface of the Grey Clay, would suggest an environment non-prohibitive to life. It is likely that the pH range within the soils, was well within normal freshwater levels, probably fluctuating between pH6-9. A higher alkalinity is precluded by the high moisture content within the soil at most times during maturation and the abundant traces of organism activity.

The final evidence for the chemical leaching effecting the formation of the Hornsleasow soils, comes from the elemental oxide and molecular weathering ratio analyses. Firstly, it can be seen that except for the high relative abundance of calcium, there is a dearth of cations (Mg^{2+}, Na^{+}, K^{+}) within the soils (Table 5.5; Fig. 5.20). The high proportion of calcium within the profiles occurs because of carbonate material included within the whole rock samples, and increases from a low of around 6% in the middle horizons of the Green and Grey Clays, to a high of up to 70% in the mixed carbonate sand and clay 'BC' regolith horizon (Fig. 5.20). The increase down the section, is accompanied by a corresponding decrease in all other oxides (sesquioxides, aluminium and iron, and the bases), except manganese. The decrease is explained by the depletion in the amount and nature of the clay minerals, the Grey Clay having high amounts of the base-poor kaolinitic clay. The manganese trend appears to closely follow the Ca^{2+} increase and can be explained accordingly. The trace element is more easily reduced to $Mn^{2+}_{(aq)}$ in organic rich acidic solutions, where it can be leached quickly out of the soil profile (Graham *et al.*, 1988). Therefore, small amounts leached from the surface might accumulate further down the profile in the more alkaline conditions of the carbonate-rich lower soil horizons.

Potassium has the highest abundance of the alkalis within the soil profiles (Table 5.5) and is seen to increase toward the Green Clay and specifically the surface horizon. Similar trends have been observed in rooted paleosols (Retallack, 1985, Freakes & Retallack, 1988, Arndorff, 1993) and are closely associated with organic cationic leaching. During the growing season, plant roots extract K^{+} from clay minerals by organic hydrolysis (Russell, 1988). Then in periods of non-growth or following the decay of the plant matter, K^{+} is released and may become relatively enriched in the root horizon. Parfenova & Yaribova (1962) have suggested that such concentrations may assist the process of illitisation within a soil (section 5.11). Therefore, the increasing proportion of K^{+} within the Green Clay may reflect the higher abundance of illite within this clay.

The other bases remain fairly low within the profile, with about 3-4% (including K^{+}) in the Green Clay and only around 2-3% in the lower paleosol. The dearth of exchangeable cations and higher proportion of Al^{3+} (Fig. 5.20) would suggest that both clays had undergone prolonged acidic weathering, with the upper clays being slightly less weathered than the lower carbonate-rich horizons! This is in direct contrast to the amount

Sample	Molecular Weathering Ratios								
	Hydrolysis			Oxidation		Hydration	Salinisation		
	silica/ alumina	bases/ alumina	silica/ bases	iron/ alumina	iron+Mn/ alumina	silica/ s.oxides	alkalis/ alumina	soda/ potash	soda/ alumina
S9 A1	5.777	1.037	6.847	0.119	0.120	5.164	0.194	0.038	0.007
S9 A2	8.645	1.880	5.123	0.119	0.121	7.723	0.192	0.047	0.009
S9 A3	7.023	2.013	3.738	0.109	0.110	6.335	0.134	0.052	0.007
S9 A4	13.738	3.588	4.018	0.128	0.130	12.184	0.169	0.060	0.010
S9 'O'	8.121	1.230	7.361	0.139	0.140	7.132	0.127	0.056	0.007
S9 A5	10.995	5.342	2.123	0.134	0.138	9.700	0.164	0.071	0.011
S9 A6	12.092	3.900	3.238	0.197	0.207	10.105	0.165	0.058	0.009
S9 'C'	6.891	27.743	0.249	0.492	0.512	4.618	0.096	0.177	0.014
S10 A1	5.645	0.751	9.922	0.113	0.114	5.072	0.182	0.039	0.007
S10 A2	5.513	0.796	9.276	0.131	0.132	4.876	0.202	0.039	0.008
S10 A3	5.510	0.497	17.724	0.116	0.117	4.935	0.186	0.036	0.006
S10 A3'	5.870	0.800	9.866	0.125	0.126	5.216	0.205	0.040	0.008
S10 A4	8.841	2.465	3.801	0.098	0.099	8.053	0.139	0.060	0.008
S10 A4'	9.183	2.317	4.223	0.105	0.108	8.314	0.142	0.060	0.008
S10 A5	9.392	2.847	3.495	0.140	0.143	8.237	0.160	0.066	0.010
S10 A6	10.324	8.485	1.242	0.132	0.146	9.117	0.171	0.063	0.010
S10 'C'	4.970	15.372	0.326	0.171	0.188	4.244	0.113	0.082	0.009
S11 A1	5.376	0.541	15.371	0.157	0.159	4.647	0.191	0.040	0.007
S11 A2	5.476	0.637	12.028	0.124	0.125	4.871	0.182	0.038	0.007
S11 A3	6.502	1.323	5.853	0.148	0.149	5.664	0.213	0.039	0.008
S11 A4	10.032	2.725	3.906	0.220	0.222	8.223	0.157	0.061	0.009
S11 A5	9.954	2.693	3.902	0.115	0.118	8.929	0.141	0.066	0.009
S11 A6	7.481	4.718	1.626	0.116	0.120	6.705	0.118	0.063	0.007
S11 'C'	9.472	19.666	0.485	0.678	0.692	5.646	0.146	0.123	0.016
S12 A1	5.407	0.589	13.368	0.169	0.170	4.623	0.184	0.037	0.007
S12 A2	6.414	0.513	20.432	0.131	0.132	5.670	0.199	0.038	0.007
S12 A3	6.468	0.604	16.100	0.134	0.136	5.705	0.202	0.038	0.007
S12 A3'	6.720	0.563	18.378	0.135	0.136	5.923	0.197	0.040	0.008
S12 A4	6.806	0.689	13.651	0.134	0.136	5.999	0.190	0.037	0.007
S12 A4'	6.765	0.583	15.659	0.129	0.130	5.993	0.151	0.046	0.007
S12 A5	7.688	0.956	9.155	0.144	0.145	6.722	0.117	0.058	0.006
S12 A6	10.028	2.181	4.910	0.220	0.222	8.221	0.139	0.067	0.009
S12 'C'	38.576	125.091	0.308	1.153	1.189	17.914	0.032		0.032

Table 5.6. Molecular weathering ratios for the four soil profiles.

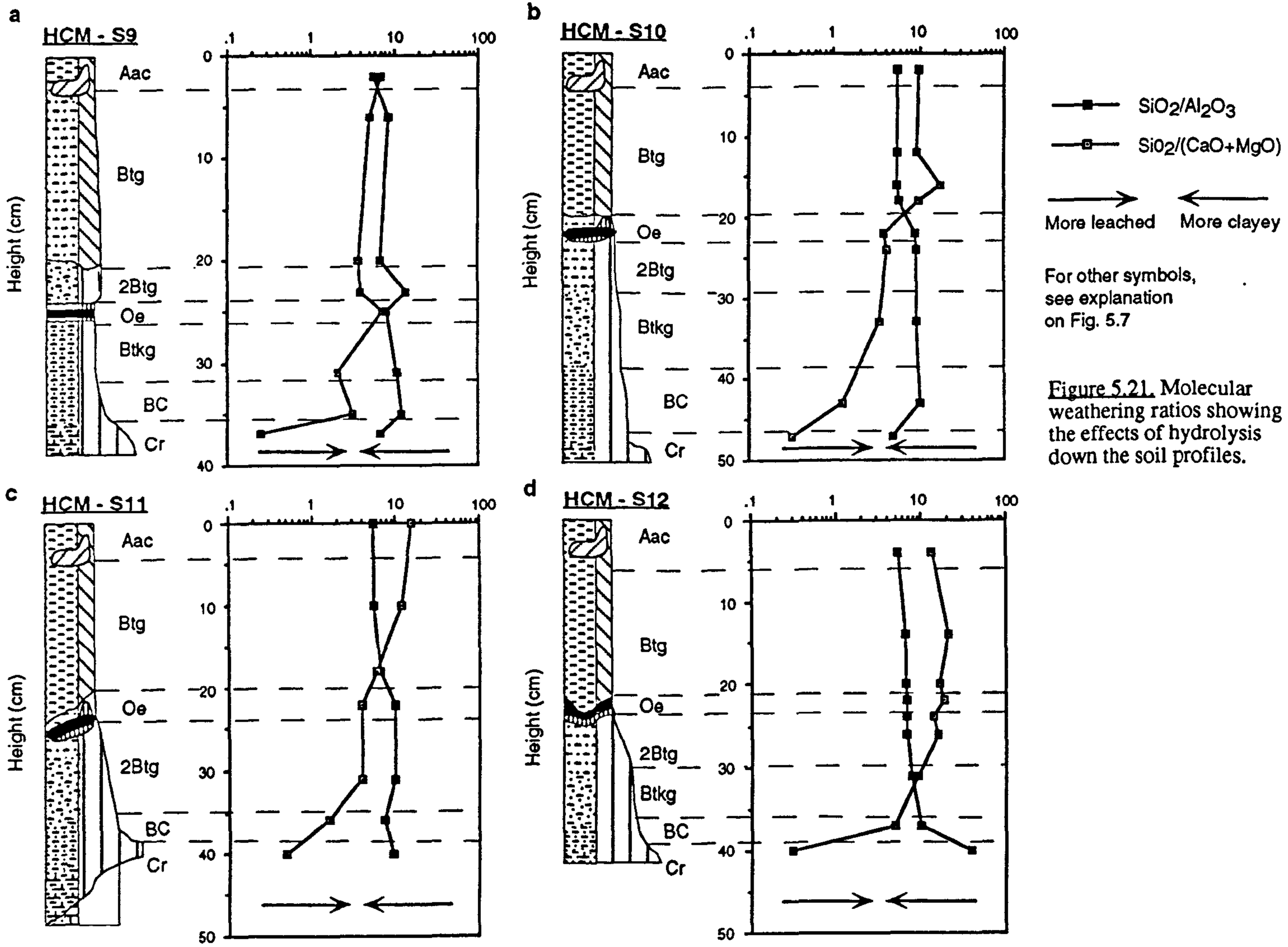


Figure 5.21. Molecular weathering ratios showing the effects of hydrolysis down the soil profiles.

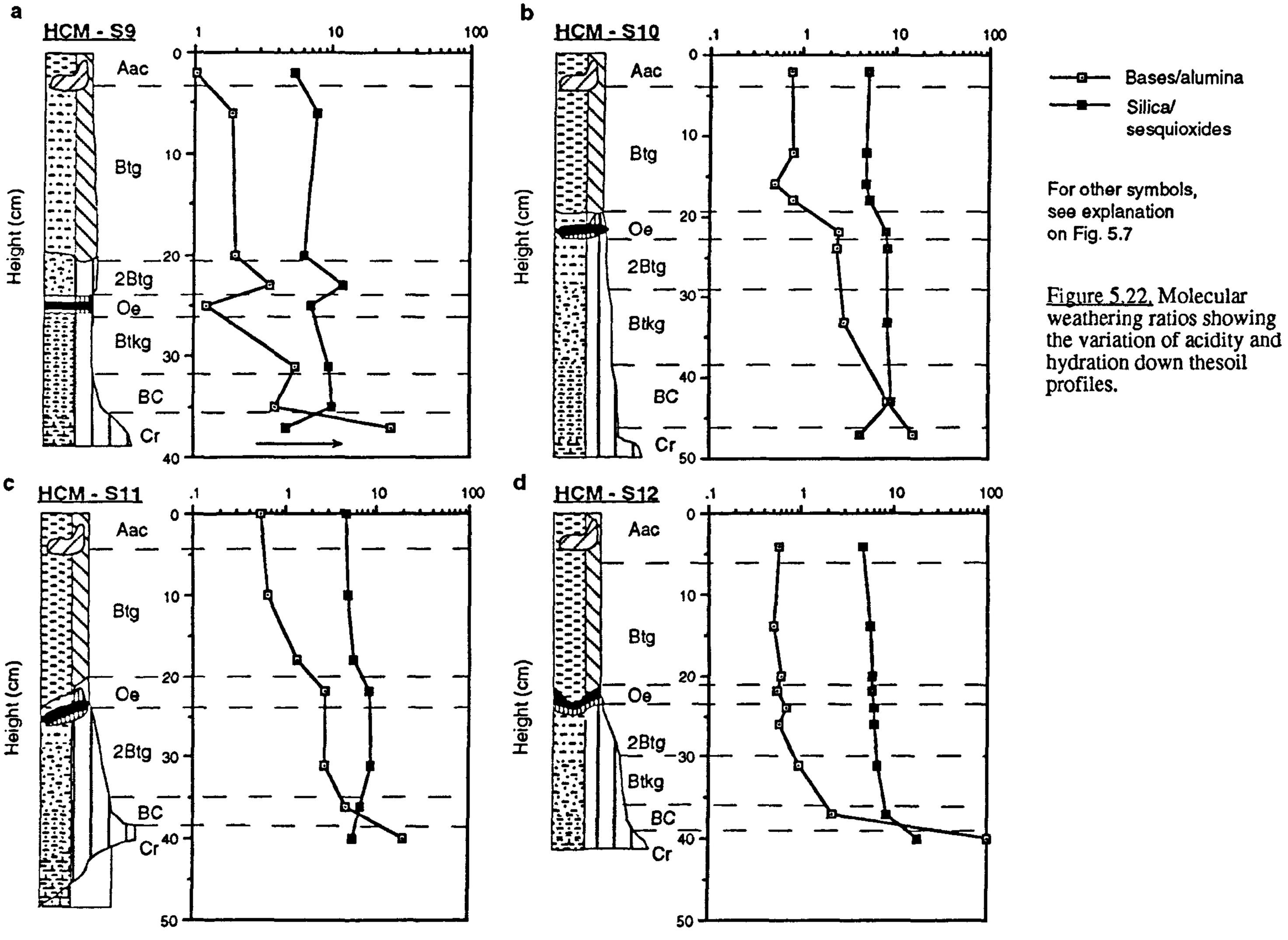


Figure 5.22. Molecular weathering ratios showing the variation of acidity and hydration down thesoil profiles.

of CaO and correspondingly $CaCO_3$ in the profiles, which suggests that pH would increase slightly down profile.

The molecular weathering ratios for estimation of hydrolysis and leaching, largely rely on measurements of the exchangeable bases compared to Al_2O_3 and SiO_2 abundances (Table 5.6) and these are affected by the anamolously high proportion of Ca^{2+} in the soils (Figs. 5.21 & 5.22; Foos, 1991). The silica/magnesium + calcium ratio shows the hydrolytic leaching patterns of clay minerals (Russell, 1988; Fig. 5.21) and this suggests an increase in the Green Clay and 'O' horizon of the Grey Clay, which is expected because of the significantly lower amounts of free Ca^{2+} in these horizons. Similarly the bases/alumina ratio (Fig. 5.22) is reliant upon measuring CaO (as well as the other free basic ions; Retallack, 1990), and therefore in most parts of the soil is greater than unity (Table 5.6) - a figure which is supposed to reflect a weakly developed or highly alkaline soil. Clearly, the soils saturated in calcium carbonate, such as the 'C' horizon, are alkaline. Interestingly, the bases/alumina calculation does show ratios as low as 0.45 within the Green Clay and the 'O' horizon of the Grey Clay (Table 5.6) and these results indicate that the soil matrix itself was undersaturated with respect to easily. hydrolysed bases, suggesting leaching in a relatively acidic environment!

To check whether the clays had suffered substantial hydrolysis in the pedogenic environment, a final molecular weathering test was applied to the data. Within highly acidic areas (pH < 4.5) clay rich in alumina is leached out of the soil profile leaving a silica rich matrix, therefore the silica/alumina ratio can help to elucidate whether conditions have been acidic enough to hydrolyse Al_2O_3 (Retallack, 1990; Foos, 1991). The values for the soil horizons showing low exchangeable bases at Hornsleasow, have relatively low SiO_2/Al_2O_3 ratios (between 9-20; Table 5.6) which fall well within near neutral to slightly alkaline soil ranges (SiO_2/Al_2O_3 ratios of between 7-16 for alkaline soils and between 16-50 for neutral soils, Retallack, 1990). The general trend of this ratio also shows the increasing clayey constituent of the soil matrix within the Green Clay soil unit (Fig. 5.22). As no large-scale quartz grain dissolution was observed, the loss of silica resulting in the low SiO_2/Al_2O_3 rations in the soils is probably from the breakdown of clay minerals and feldspar grains, rather than hydrolysis of quartz (cf. Foos, 1991).

In conclusion, although the molecular weathering ratios suggest some sort of removal of bases within the soil weathering profile, all the other evidence points to fairly neutral or slightly alkaline palaeoenvironment, and another source of leaching must be proposed (section 5.11). The pH range for the Hornsleasow paleosols was probably within the region of pH7-9, although in localised microenvironments, near to decaying organic matter or within the root zone, more acidic levels might have evolved (pH5-7).

(b) *Oxidation, reduction and soil Eh.* Oxidation reactions are those in which an element loses an electron to form a cation. The reverse reaction involving

electron gain, is called reduction. The most common reversible reaction involving oxidation and reduction is the conversion of Fe^{3+} to Fe^{2+} and vice versa. The electron potential (Eh/V) of a soil indicates the oxidation status of the horizon. A positive Eh is found within oxidising them, whilst a negative Eh is strongly suggestive of a reducing environment (Fig. 5.18). The principal oxidising agent in modern soils is oxygen and its exclusion from soils is largely a consequence of waterlogging conditions. Baas-Becking *et al.* (1960) suggested three main categories of redox potential within soils: reducing or waterlogged soils; intermediate or periodically wet soils; and oxidising or well drained soils. Eh can be directly measured in modern soils, but must be inferred from studies of mineral associations, organic matter preservation and chemical weathering profiles in paleosol horizons (Retallack, 1990).

Although carbonate and phosphatic fossils are not affected by differences in Eh conditions, the preservation potential of plant tissues is greatly enhanced by the reduction of aerobic microbial decomposers within waterlogged soils. The exceptional preservation of the tree resin, spore and pollen cuticle and coalified plant remains within the lower Grey Clay horizons (section 8.3.3) indicate a rather reducing gleyed environment for this unit. The best preservation of the plant material is in the basal layer, and this suggests that this at least was permanently waterlogged. The ped structure (section 5.6) and dull or weak anisotropic microfabric (section 5.7) of the deeper Grey Clay soil horizons also suggests a poorly-developed gleyed soil, with little or no active reworking of the soil matrix by animal burrowing. The surface 'O' horizon contains the quartz-lined burrows, which suggest invertebrate activity within a wet soil.

The greenish-grey colour of the Grey Clay is overwhelming evidence that reducing conditions prevailed within the unit and the presence of localised pyritic fossils also suggests Eh conditions as low as 0.3V (Fig. 5.12 & 5.18). However much of this pyritisation is associated with decaying organic matter and locally increased acidity levels. The lower pH in these regions facilitates pyrite growth even in relatively 'oxidising' regions of the paleosol (Eh of around 0V at pH 4-5; Fig. 5.12). The poor crystallinity of the illite and relative abundance of kaolinite within the Grey Clay actually suggests humid weathering within a comparatively aerobic soil. The apparent disparity could be caused by a fluctuating water table, or a pronounced seasonality (Retallack, 1990).

The Green Clay contains less well preserved plant debris, including rare palynofloral extracts and resin, indicating an intermediate redox potential in this unit. Vertical rootlets pervade the layer, suggesting that this layer was not permanently waterlogged. These usually occur as oxidised limonitic stains or ferric oxyhydrite concretions. The roots penetrate into the top horizons of the underlying Grey Clay, where they show characteristic limonitic staining. This mottling occurs because the lower

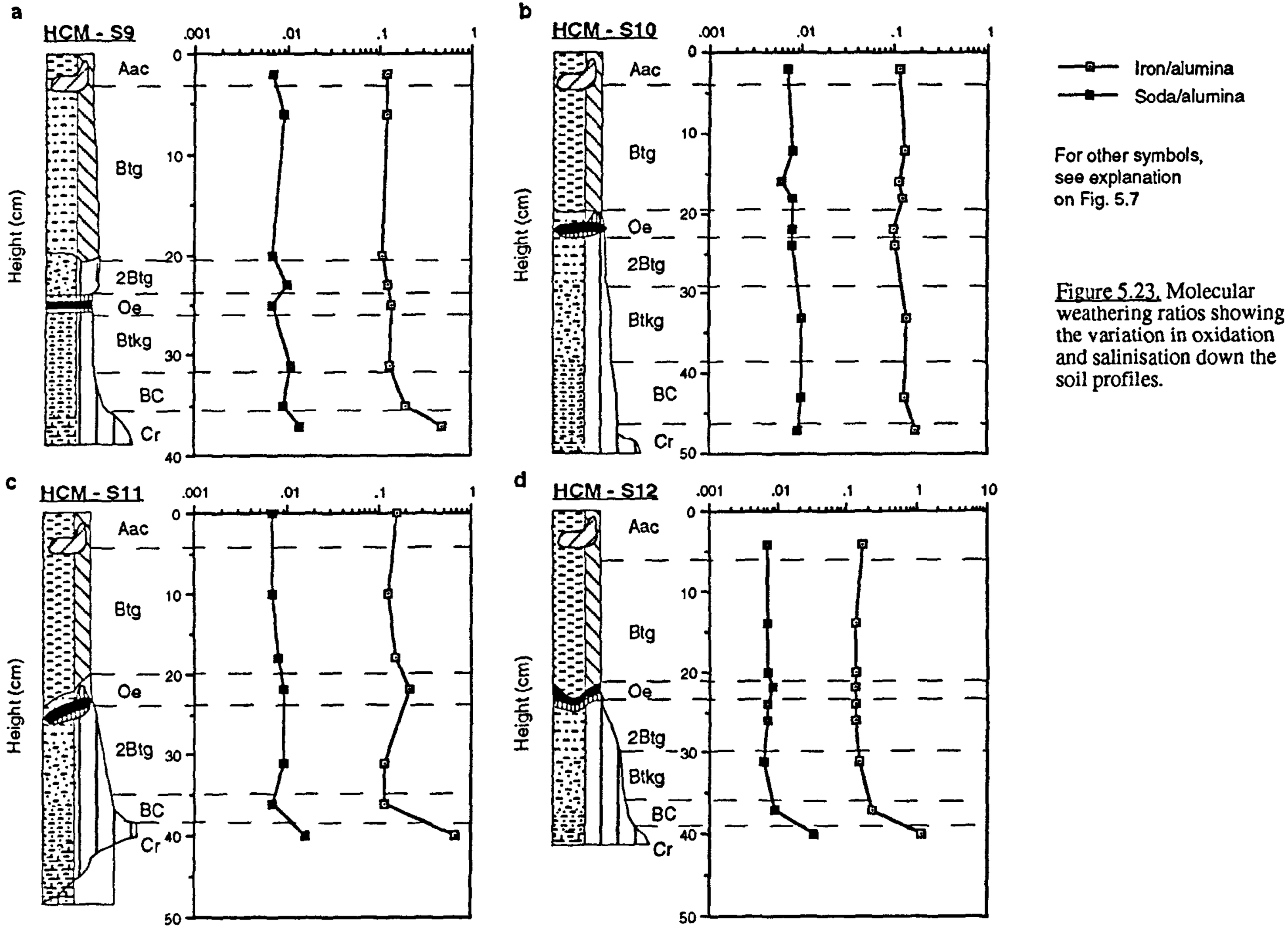
horizon was still waterlogged following the deposition and during the maturation of the Green Clay. This situation is called 'groundwater gley' (Retallack, 1990).

The presence of charcoal within both clay units indicates a seasonal climate dry enough for forest fires to occur, and the blocky ped structure and anisotropic microfabric developed within the Green Clay indicate that this clay may have been subject to the fluctuating precipitation levels of a seasonal climate. Iron mottling within the Green Clay is also highly suggestive of pronounced seasonality, with hydrolysis reactions and root growth during wet periods liberating ferrous iron from soil minerals. During the dry season this is oxidised and fixed as goethite or limonite to form nodules and rhizocretions. There is no development of an organic peaty horizon at the surface of the Green Clay, which also suggests that the surface horizon was fairly well-drained (Retallack, 1990). Glaebules such as siderite and carbonate concretions and nodules are also characteristic of soils with intermediate redox potentials.

The chemical profiles and molecular weathering reactions of the soils can also be useful to quantify soil redox potential. For example, the dominance of iron over manganese within the soil weathering profiles (Table 5.5; Fig. 5.20) and presumably Fe^{3+} over Fe^{2+} (as indicated by the abundance of ferric compounds within the soils) reflects a high oxidation potential within them. Total iron falls off within the lower Grey Clay horizons and is accompanied by a rise in MnO_2 (Fig. 5.20), which suggests that the oxidation levels were lower within the gleyed unit. Ferric/ferrous iron ratios are the most effective measure of redox potential within a paleosol analysis, but the AAS is not sufficiently sensitive to measure the individual atomic abundance. Therefore as a rough guide the total iron and total iron plus manganese levels were assessed against alumina (Table 5.6). These ratios tend to be close to zero in reducing or intermediate environments and between 1-1.5 in oxic horizons (Retallack, 1990). The profiles (Fig. 5.23) indicate that the ratios are low within all clay horizons (<0.4), although significantly, they tend to increase toward the top horizon ('O' and 'A') of both units.

The lowest part of the soil profile, the deeply weathered *terra-rossa* rich haematitic deposits and mixed carbonate regolith soils overlying the palaeokarstic surface are extremely well drained (Fig. 5.12 & 5.18). These exhibit high total iron/alumina and total iron plus MnO_2/alumina ratios (>1), suggesting that these where formed within an oxic soil environment.

(c) *Hydration, dehydration, salinisation and soil moisture content.* Hydration or dehydration reactions involve the addition or loss of water which is structurally part of a mineral lattice. Common hydrated minerals in the Hornsleasow soils are the ferrihydrites, such as goethite and limonite, which can lose interstitial water to produce haematite. The degree of hydration of a soil is essentially the same as the



For other symbols, see explanation on Fig. 5.7

Figure 5.23. Molecular weathering ratios showing the variation in oxidation and salinisation down the soil profiles.

degree of waterlogging and therefore many of the arguments put forward to support the redox potential of the Hornsleasow soils also apply here.

Molecular weathering ratios give a crude measure of former hydration in paleosols, for example the ratio silica/sesquioxides, is relatively high (>10) in quartz rich soils with few hydrated minerals, and conversely low (<10) in ferruginous, clayey soils, which contain abundant hydrated clay and iron minerals (Retallack, 1990). The weathering profile for the Hornsleasow soil units (Table 5.6; Fig. 5.22), indicates that the minerals are largely in a hydrated phase. Minerals can also be hydrated or dehydrated during diagenesis, for instance the illitisation of smectite and the formation of haematite are both common during deep burial of paleosol horizons (section 5.11).

Salinisation only occurs within truly arid climates, where soluble salts such as gypsum and halite precipitate within the soil profile. Calcrete nodules can also be suggestive of an arid climate, but also occur within more humid weathering regimes in limestone or coastal regions. Salinisation products are not present within the Hornsleasow paleosol profile and this is also substantiated by the low sodic to alumina ratios plotted on Fig. 5.23.

5.11. Diagenesis and alteration of the soils after burial

Many of the pedogenic processes which take place in the subsurface horizons of soils, such as chemical weathering and glaebule formation, are "diagenetic" alteration of the original sediment. Therefore, in considering the true diagenetic changes which have modified the paleosol horizons after final burial, the study must differentiate between the changes which modify a sediment during soil formation from those of later diagenetic and metamorphic alteration. The distinction is a fine one, and can be made only from detailed field and laboratory studies of soil sections and profiles.

The relationship between a feature which is suspected of being either pedogenic or diagenetic in origin, such as a nodule or concretion, and a definite soil feature, such as root traces, should be established in most circumstances. The general appearance of these features may help to elucidate their origin, for instance ferruginous, phosphatic and pyritous nodules which have formed around or nucleated within unsquashed bioclasts, plant fragments or bones suggest early replacement within the soil regime. Wholesale pyritisation of woody tissue is largely a late diagenetic feature (Malton, 1973), but this does not occur within the Hornsleasow unit.

A general rule for establishing diagenetic history of a fossil soil profile is to look for features incongruous within the suggested palaeoenvironment of the soil, although this can also occur within a soil developing within a rapidly changing environment or a exhumed buried profile. For example, if the Hornsleasow ferruginous nodules and rhizocretions were heavily oxidised to haematite, this would suggest that diagenetic

dehydration had taken place (Walker, 1967), as the palaeoenvironment for both soils indicates a fairly reduced redox potential. Burial gleying can also occur within buried waterlogged soils rich in organic matter. The green coloration of the Green Clay was probably caused by this reaction, as the other features of the unit do not suggest a permanently waterlogged horizon. Although the MgO enrichment in the upper Green Clay compared with the lower Grey Clay and *terra-rossa* soils seems to follow the other base trends in the weathering profile (Fig. 5.20), it might also suggest a degree of marine water alteration of the upper clay following the transgression in later Early Bathonian times (section 5.13; Arndorff, 1993).

This enrichment might also support the earlier claim that smectitic and particularly chlorite clays (both high in Mg^{2+}) are present in the Green Clay (Vaughan, 1989). Burial increases the loss of cations such as Mg^{2+} and Na^{+} and a common diagenetic reaction is the alteration of smectitic clays to illite by illitisation. This reaction is accompanied by the reduction of potash-rich minerals such as the feldspars (rare in Hornsleasow aggregates) and is promoted by the removal of other common cations (Na^{+}, Ca^{2+}, Mg^{2+} and Si^{2+}) within the soil pore waters (Retallack, 1990), at depths of 2000-3000m and temperatures of between 60 and 180°C (Curtis, 1985). Although, theoretically illitisation could occur at much lower temperatures and depths over long periods of time (Retallack, 1990). The poor crystallinity of the illite within the Grey Clay suggests that it is an alteration product (Frey, 1987). Future determination of the clay mineral composition will be crucial in ascertaining the true diagenetic history of the two clay units.

5.11.1. Compaction and cementation

Compaction and cementation of soil horizons can take place even within the lower levels of an active soil profile. Following the deposition of overburden these processes include dewatering, reduction in volume and finally, lithification of paleosol horizons. In the case of the Hornsleasow paleosols, the non-lithified and waterlogged state of the clays prior to excavation suggests only minor modification by compaction. The near vertical root traces of the Green Clay also indicate that the clays were not significantly compacted following burial. However, plant debris is severely flattened in most cases and some ferruginous glaebules, particularly those nucleated around freshwater gastropod remains, appear squashed prior to induration. Much of this flattening could have happened within the lower soil horizons following deposition of the overburden (cf. Elliot, 1985).

Most of the vertebrate material is broken (section 10.4), but very little of this can be explained solely by compaction fragmentation. However, the large cetiosaur bones which occur at the base of the hollow do show large centimetre cracks which may have up to five millimetres shear displacement indicating substantial downward force. These

shattered bones often display a diagenetic sparry calcite cement occurring within these cracks, which may have been precipitated during late burial.

Slight compaction of the clays is indicated by the occurrence of slickenslide microfabrics which cut through earlier texture within thin sections taken from the paleosol units. These occur because of slight, local differences in geochemistry or mineralogy of the clay matrix causing preferential shearing during compaction. For instance, clays enriched with calcium and potassium are not so easily compacted as those rich in magnesium, sodium and iron. Also the structure of kaolinite means that it is much less compressible than smectitic clay (Retallack, 1990). The presence of abundant organic matter or silt-sized particles within the matrix can also affect the compressibility of clays.

5.12. Interpretation and classification of soil types

Modern soils are categorised and classified according to three main systems. These work in a similar hierarchical manner that taxonomy provides for plants and animals. The three modern schemes most commonly employed in soil science are: (1) *The handbook of Australian soils*, published by the Commonwealth Scientific and Industrial Organisation (CSIRO) of Australia (Stace *et al.*, 1968); (2) *The Soil Map* of the Food and Agricultural Organisation (FAO) of UNESCO (1971-1981); and (3) *Soil taxonomy of the Soil Conservation Service*, published by the Soil Survey Staff of the United States Department of Agriculture (USDA, 1975).

However, there are several problems associated with the use of these schemes in order to classify fossil soils. The primary problem is that many of the features on which the systems are established do not preserve well within ancient paleosol sequences. Colour, moisture content, bulk density and some chemical parameters, such as cation exchange capacity, are easily altered during burial and diagenesis (Yaalon, 1971). Although in some cases, an approximate determination of these features can be made, a quantitative analysis cannot be made. Secondly, many soil classification schemes rely upon a detailed knowledge of the prevailing climatic conditions in their determination of a soil type. For instance, *The Soil Map* is particularly useful for classifying tropical soils, whilst the *Soil taxonomy of the Soil Conservation Service* works well for soils developed within temperate climates. The soil type 'Aridosol' defined by the latter scheme, can only be recognised if an estimation of the regional precipitation levels and temperature can be made over a period of 90 days! It would be almost impossible to find this type of information within a fossilised soil profile.

In order to classify ancient soils, Mack *et al.* (1993) presented a simplified classification system based upon pedogenic features which have the greatest chance of being preserved and recognised within a paleosol horizon. The six features which they define as most important are : organic matter content and preservation; development of

pedogenic horizons or 'horizonation'; redox conditions; *in situ* mineral alteration; illuviation of insoluble minerals; and accumulation of soluble minerals (Mack *et al.*, 1993). The scheme is hierarchical but extremely simple and flexible, as the most prominent feature recognised within the paleosol determines the paleosol order or type, with subordinate features appearing as modifiers prefixing the order name. The nine paleosol orders are : (i) *Histosol* (coal seams, with more than 70% by volume carbonaceous material); (ii) *Spodosol* (paleosols with strongly developed 'spodic' horizons derived from illuviation, for example illuvial buried organic matter or iron oxides); (iii) *Oxisol* (paleosols which show evidence of severe chemical alteration of chemically unstable minerals to clay and sesquioxides, i.e. the development of an 'oxic' horizon); (iv) *Vertisol* (paleosols which exhibit homogenisation of the profile by pedoturbation processes such as shrinking and swelling of expandable clay minerals, i.e. with a well formed 'vertic' horizon); (v) *Calcisol* (paleosols which exhibit a subsurface 'calcic' horizon which is enriched in pedogenic carbonate material); (vi) *Gypsisol* (paleosols displaying a surface or subsurface 'gypsic' horizon enriched in pedogenic gypsum or anhydrite); (vii) *Gleysol* (paleosols which have surface or subsurface gleyed horizons with low redox conditions); (viii) *Argillosol* (paleosols containing a prominent subsurface accumulation of illuvial clay known as an 'argillic' horizon); and (ix) *Protosol* (weakly developed paleosols with little pedogenic horizonation, sedimentary textures still discernible). The first four types are taken from the *Soil taxonomy of the Soil Conservation Service*, although with modification to the criteria for recognition, whilst the rest are defined in detail by Mack *et al.* (1993). They also define 18 subordinate modifiers which can be applied to the general soil orders. Such subordinate features include self-explanatory prefixes such as carbonaceous, calcic, argillic, concretionary, ferric, gleyed, and nodular (Mack *et al.*, 1993).

Mack *et al.* (1993) also identified three significant problems which can arise when trying to categorise paleosol units in such terms. The first is the effect of concomitant deposition of sediment and pedogenesis which would give rise to a cumulate soil profile, exhibiting pedogenic features resulting from several periods of soil formation. The second factor occurs within an evolving pedogenic regime, for example a buried soil profile being exhumed and re-developed, or a soil developing under rapidly changing climatic controls and resulting in a polygenetic paleosol profile. Again this would complicate soil recognition by superimposing features formed through different pedogenic effects. The third problem arises from erosion of a soil profile prior to final burial and fossilisation. Clearly soil truncation could potentially remove key horizons from a paleosol profile. Although, the Mack *et al.* classification of paleosols has not been accepted by mainstream palaeopedology, in the light of the factors outlined above and based upon the results set out in the preceding sections, the Hornsleasow clay units are

classified below in terms of the scheme by Mack *et al.* (1993). In terms of the older classifications, it is acceptable to term both paleosol horizons at Hornsleasow, Entisols (V.P. Wright, pers. comm.).

The Hornsleasow paleosols are also categorised by the type and relationships of the fossil vegetation which they contain as fossil plant formations. This is based upon rough guidelines set out by Retallack (1992) upon how to classify fossil plant-bearing terrestrial sediments (including paleosols) in terms of their palaeoecology. This included a review of the types of modern and ancient plant-bearing environments to produce a comprehensive classification. The diagnostic features recognised for plant formation types include the overall soil profile, root trace depth and a rough appraisal of the organic productivity and biomass, in terms of organic matter content, with allowances made for diagenetic alteration (Retallack, 1990, 1991). Fossil plants are best preserved within poorly drained soils, although caliche and ferruginous rhizocretions may preserve an indication of former vegetation in better drained conditions.

The time taken for a soil profile to form varies from days (for some Podsol horizons) to millions of years (Retallack, 1990). Superposition is not easily resolved for a single paleosol unit, because of the constant flux involved in soil maturation. Few features in soil profiles vary linearly with age, and hence, establishing an estimate for the maturation time is extremely difficult, and this is exacerbated in a paleosol because of the the effects of erosion and later diagenetic alteration. A general qualitative scale for soil development has been suggested by Retallack (1990), based upon the varying degrees of formation of calcic horizons, soil microstructure and peat accumulation.

5.12.1. Grey Clay soil formation and classification

The clay paleosol units overlying the *terra rossa* horizon at the base of the limestone hollow, represent a cumulate soil profile, with features of the underlying Grey Clay being overprinted by Green Clay deposition and pedogenesis. The deposition of the Grey Clay, encouraged waterlogging within the hollow, and the most obvious pedogenic feature of the Grey Clay is the gleyed redox conditions. Therefore in terms of the scheme devised by Mack *et al.* (1993), the Grey Clay is a Gleysol. Subordinate modifiers to this classification are the large amounts of unoxidised plant matter which the unit contains, the calcareous nodules and the clayey nature of the matrix. Taking account of these factors the Grey Clay is a Carbonaceous Calcic Argillic Gleysol.

The presence of a truncated drab grey and black 'O' horizon at the top of the Grey Clay and well-preserved carbonaceous matter within the unit, suggests that the soil was waterlogged. Although no root traces are discernible in the sections studied, rare rhizocretions and pyritised roots have been found and identified as lycopodaceous. Waterlogged soils bearing herbaceous flora, such as lycopods and ferns can be termed

marsh deposits (Retallack, 1992). The presence in the soil of calcareous nodules and its general position within a limestone landscape, suggests an alkaline marsh or fen-type paleosol based on Retallack's guidelines (1992). The fauna of the deposit is dominated by semi-aquatic tetrapods, with some associated ?freshwater invertebrates and aquatic vertebrates and their presence supports a marsh or pond environment. Of the 6000 fossils sieved originally, the majority came from the Grey Clay (Metcalf *et al.*, 1992).

Although no calcic horizon occurs in either of the paleosol units, the calcareous nodules and other indicators of better drainage within the Grey Clay profile, indicate that the waterlogging conditions were either transient or seasonal allowing growth of these forms, or that these features developed within the Grey Clay following deposition of the overlying Green Clay unit and the subsequent desiccation or silting up of the marshy pool (see below). The former hypothesis suggests a cumulate profile for the Grey Clay, whilst the latter is indicative of a polygenetic formation.

The clayey matrix of the Grey Clay 'B' horizons possesses a weak to moderate formed ped structure, with dull and weakly anisotropic microstructures, and the possibilty of relict bedding. This suggests that the clay paleosol took anything from 100-10 000 years to develop, taking into account that the effects of pedogenesis may have been subdued by the waterlogging. The surface accumulation of coalified peat has been compacted to around 20-50mm in depth, and suggests a minimum period of between 100-1000 years of formation (Retallack, 1990).

5.12.2. Green Clay soil formation and classification

The mineralogical and geochemical analyses suggest that the Green Clay was derived from the same source as that of the Grey Clay but following deposition was subject to less intense weathering, perhaps as a result of gleying. However, the horizon was also subjected to a fluctuating moisture content and eventually dried up at the surface, as made evident by the ferruginous rhizocretions and strong oxic mottling in the 'A' horizon of the unit. This conflicting evidence would suggest that the Green Clay soil profile is polygenetic. In terms of the strong evidence for oxidation and accumulation of kaolinite clay and sesquioxides, it could be termed an Oxisol (after Mack *et al.*, 1993). Reducing waterlogged conditions never seem to have been as important as in the underlying Grey Clay, and to suggest that the soil could be a Gleysol would not fit the evidence. However, although it would seem almost a contradiction in terminology, which highlights the problems with the Mack *et al.* hierarchical scheme, the subordinate modifier 'Gleyed' Oxisol could be suggested based upon all the available analyses. A less fantastic classification could be based solely upon the predominantly clayey matrix and this would also apply to the underlying Grey Clay, and both could be termed Argillosols. However, with the lack of unequivocal proof of the composition of the

original sediments (section 6.1), and the potential diagenetic alteration of the clay minerals, I would not endorse an illuvial origin for the clayey matrix. More preferable is the application of the subordinate modifier 'Argillic'. Therefore, in terms of the classification scheme proposed by Mack *et al* (1993), the Green Clay is best defined as a Gleyed Argillic Oxisol.

The Green Clay apparently was thickly vegetated, at least prior to its final burial by the overlying limestone, since oxidised lycopodaceous and pteridophyte plant debris was discovered on the surface of the 'A' horizon, which is pervaded by root traces and ferruginous rhizocretions of the same flora. Taking into account the polygenetic nature of the Green Clay, the soil could also be termed an alkaline marsh or fen environment, but the seasonal fluctuation and eventual desiccation of this soil profile suggests that the plant formation was better drained than that of the Grey Clay and is more properly defined as herbaceous brakeland based on the classification scheme of Retallack (1992).

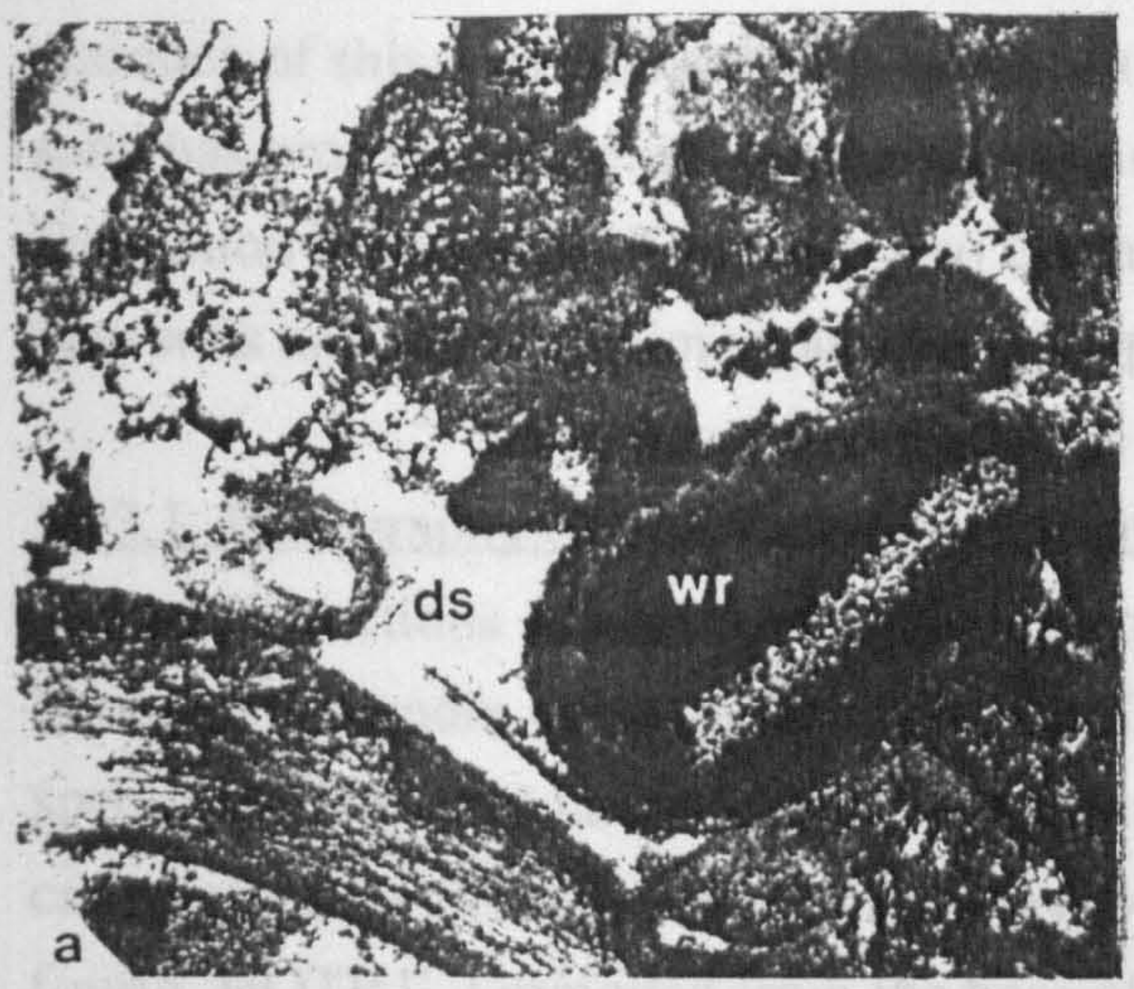
The ped-structure and microstructure of the Green Clay paleosol matrix is moderate to strong in development, and suggests a minimum time of soil formation of 10^4 - 10^5 years (Retallack, 1990). The 'A' horizon rhizosphere also support this sort of time-scale.

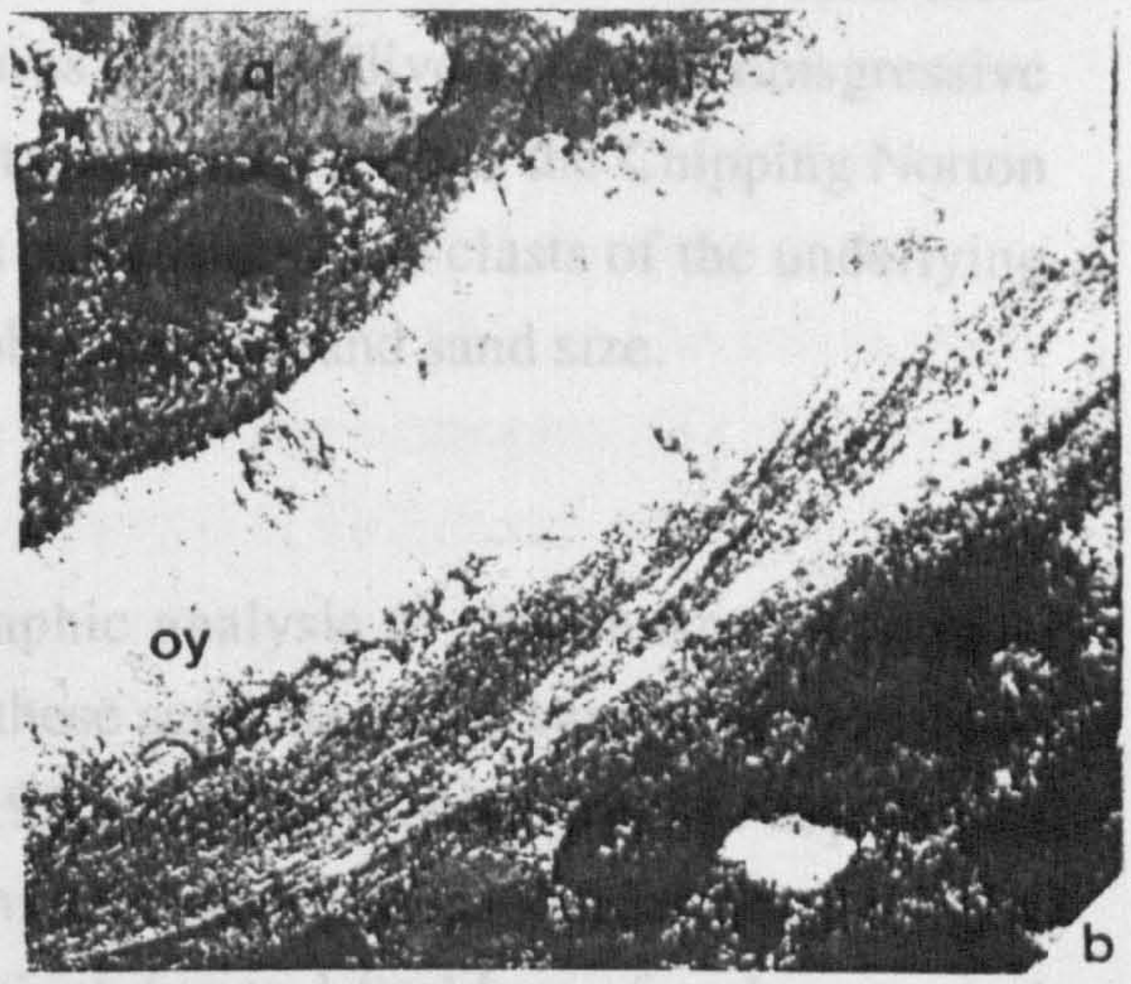
5.13. The overlying transgressive sand unit

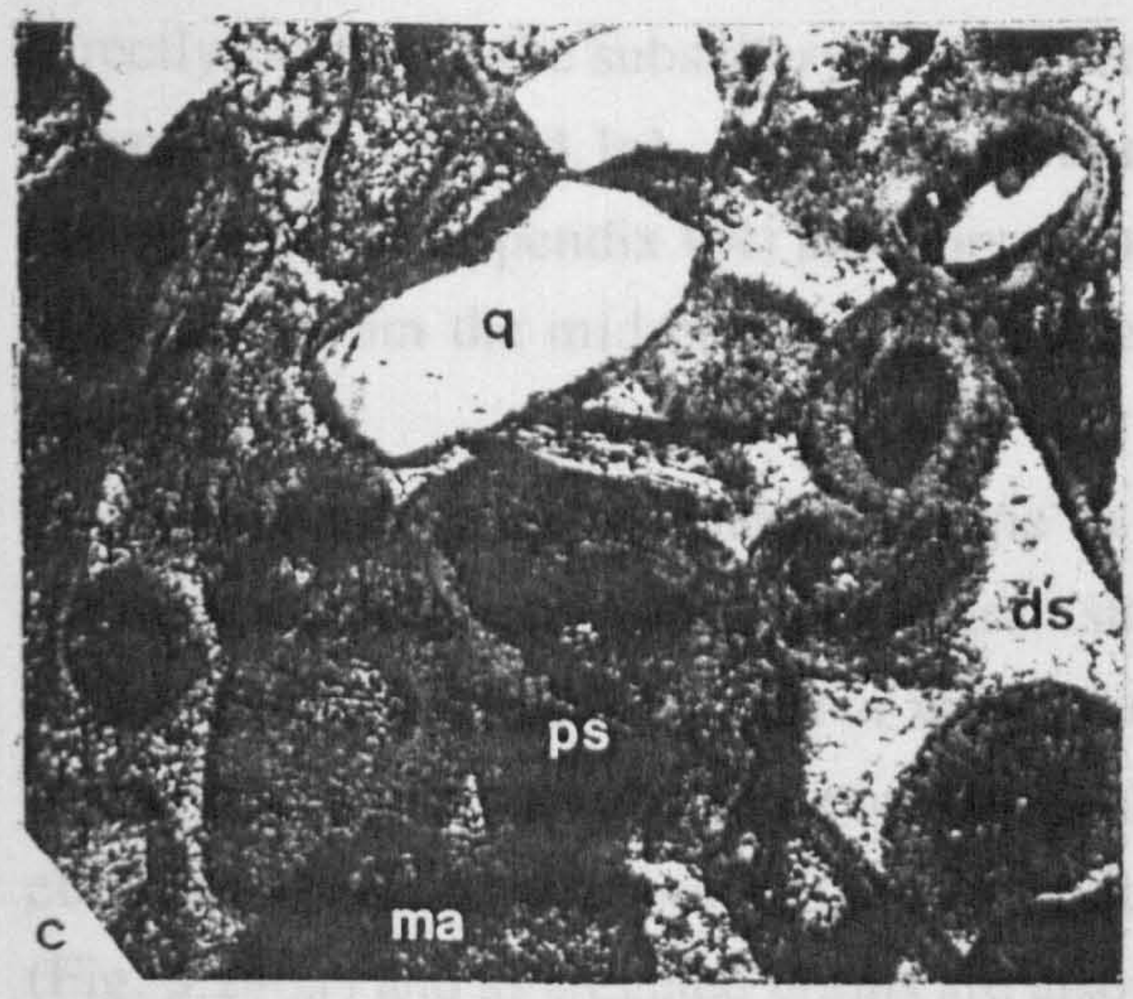
The top surface of the Green Clay paleosol at Hornsleasow is truncated and overlain by a fine-grained sandy carbonate unit, known as the 'transgressive sand layer' (Metcalf *et al.*, 1992). This horizon is 0.2-1.0 metres thick in places and can be seen to directly overlie parts of the palaeokarstic land surface from which the clays have been removed (Fig. 4.1a). The transgressive sands are overlain by a cross-bedded biooosparite unit of the Chipping Norton Formation, the base of this bed is erosive and is seen in some places to scour down into the underlying sands (Metcalf *et al.*, 1992).

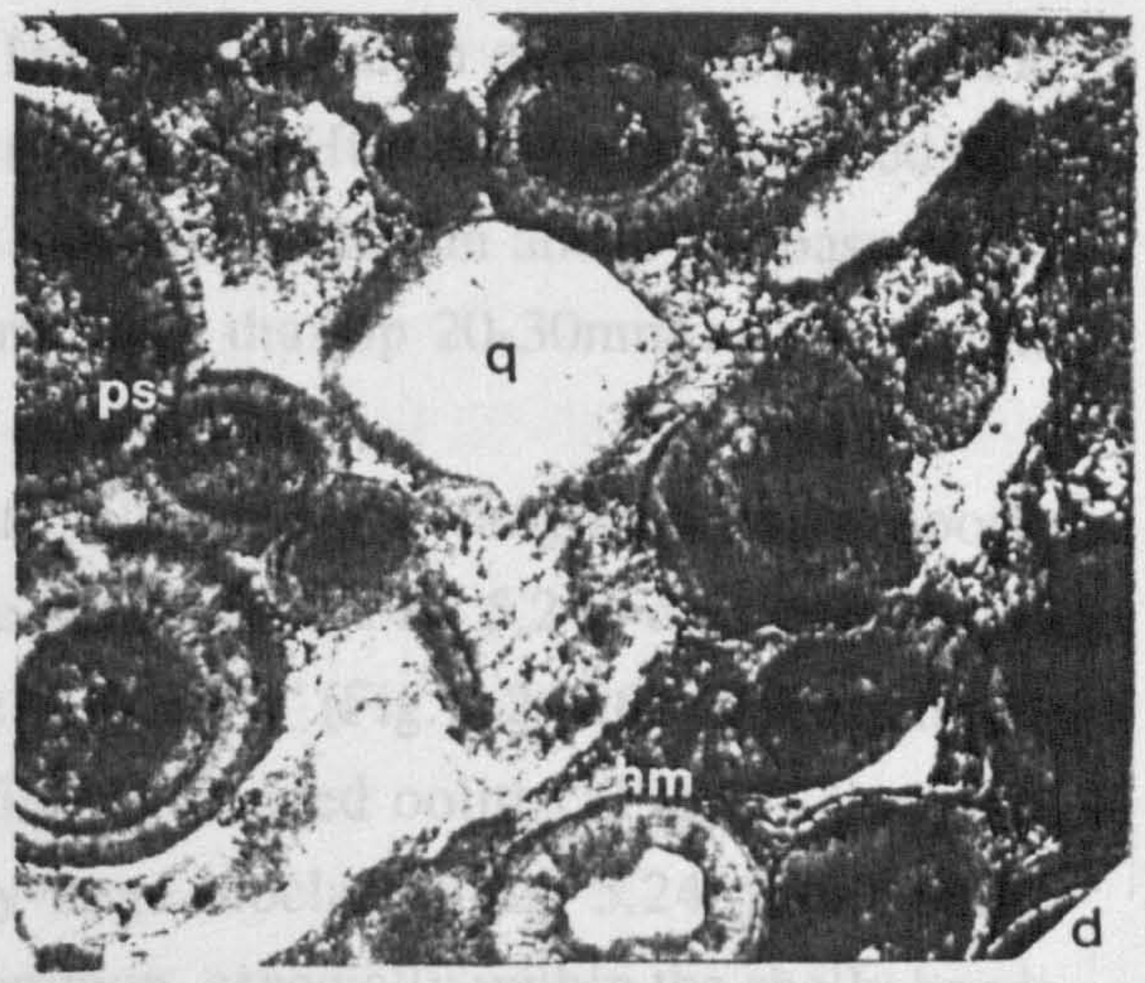
In the field and in hand specimen the appearance of the rock is of a closely packed calcarenite. In places it can be seen to be fairly well-cemented by a sparry cement, whilst in others it resembles a carbonate mud-dominated wackestone or packstone. It is distinctly planar-laminated upon a fine centimetre scale, and this lamination is clearly seen in both weathered field and hand specimens. In some places low angle cross lamination with associated bed truncations can be observed. The limestone is a pale creamy buff colour, weathering at outcrop to a pale yellow-brown friable sandy deposit, with high associated fenestral porosity developed along bedding planes.

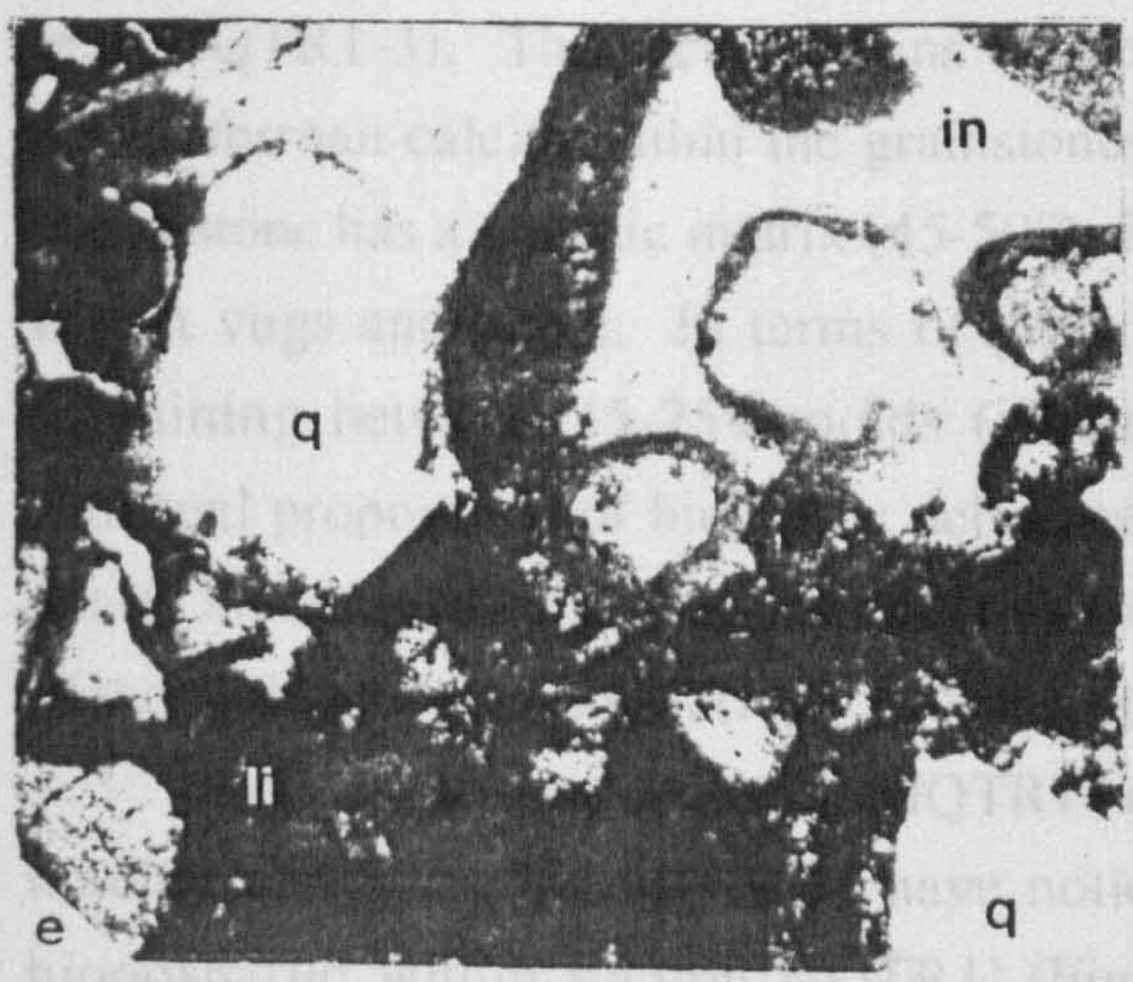
Although the rock is predominantly a well-sorted fine-sand, in places it contains thin lenticular bands of coarsely bioclastic debris. These shelly bands comprise whole and fragmental bioclasts, including recognisable oysters, pectinid and trigonid bivalves, rhynchonellid brachiopods, echinoid fragments and numerous fish teeth, including











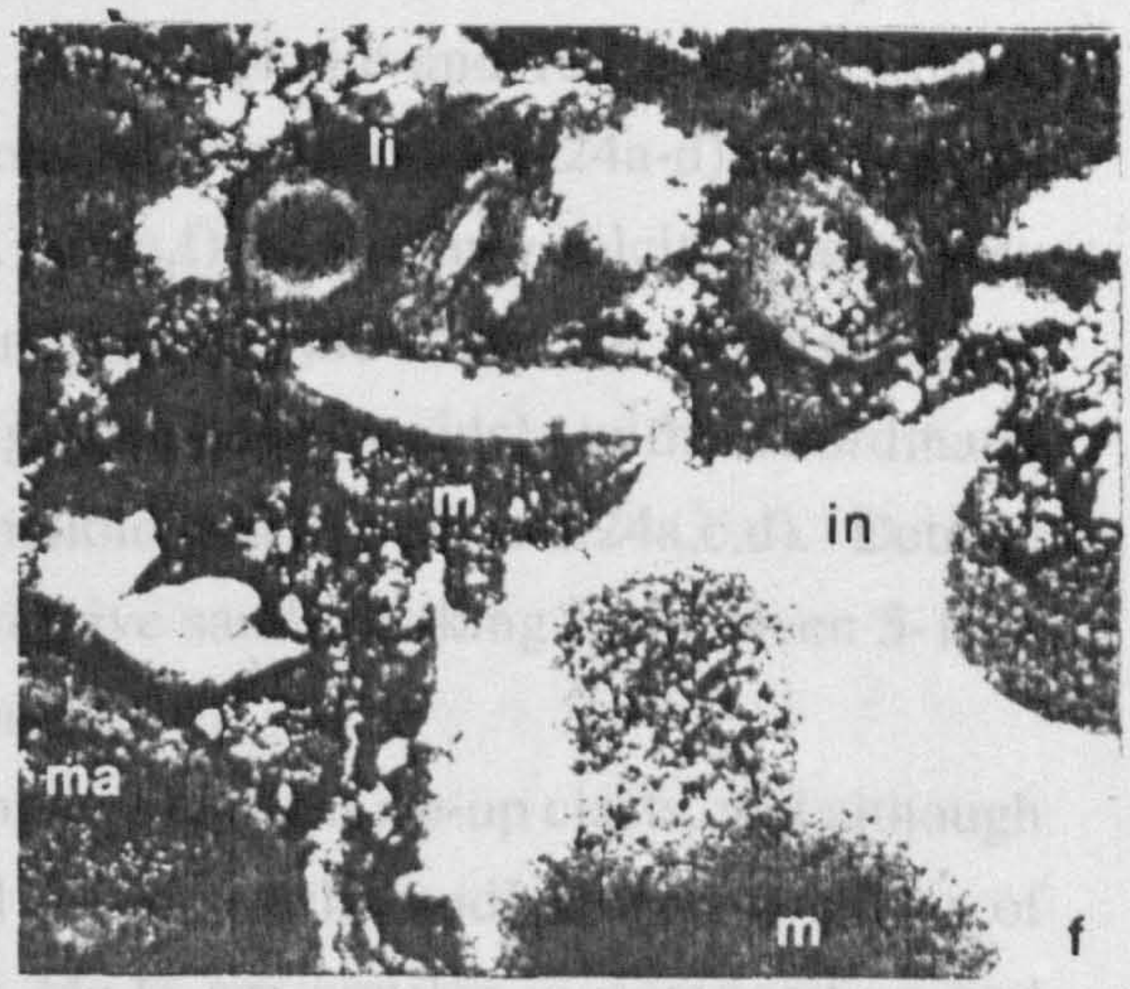


Figure 5.24. Photomicrographs of petrographic thin sections of the 'transgressive sand' unit overlying the paleosols and karst surface: (a) section HQTR1-1, cross polarised light; (b) HQTR1-1, cross polarised light; (c) HQTR1-2, plane polarised light; (d) HQTR1-3, plane polarised light; (e) HQTR1', cross polarised light; (f) HQTR1', plane polarised light. Field of view 2.3mm. All sections were photographed under blue-filtered light. Abbreviations as in Figs. 3.5 and 4.8.

pycnodontid and shark remains. Marine dinocysts have also been recovered from residues of this horizon (section 8.3.1). In terms of biotic diversity, the transgressive sand has probably the most diverse macroinvertebrate fauna within the Chipping Norton Formation limestones. The layer also contains abundant rip-up clasts of the underlying clay soils and karstified limestones of small pebble, granule and sand size.

5.13.1. Petrography of transgressive sand unit

Four thin sections were prepared for petrographic analysis of the transgressive sand horizon (Appendix C4; Figs. C1 & 5.24). Of these sections, three came from the same specimen (HQTR1) taken by M.J. Simms in 1990 from the 0.4-0.5m thick deposits of carbonate sand directly overlying the clays within the main hollow at Hornsleasow. The fourth, HQTR1', I removed from the base of a thick (up to 1.0m) lens of carbonate sand directly overlying the subsidiary clay horizon located some 20m or so west of the main excavated site (Fig. 4.1a). The three sections taken from HQTR1, were numbered one, two and three (Appendix C4) and correspond to positions 10mm above the base of the sand unit, from the mid part of the sample and from the top 20-30mm of the sample respectively.

Compositionally, the limestone is laterally variable, ranging from a poorly washed oosparite and oopelsparite (grainstone-packstones; Fig. 5.24a-d) overlying the main lens to a packed oomicrite or wackestone in HQTR1' (Fig. 5.24e,f). The grain size of the majority of the layer is composed of fine sand-sized ooliths (Fig. 5.24c,d) and could be termed micro-oolitic. Exceptionally large bioclasts (Fig. 5.24b), intraclasts (Fig. 5.24e,f) and aggregated grains are also common, especially within the shelly bands (e.g. HQTR1-3). The rock contains between 5-20% micrite and 15-25% pore-filling sparry ferroan calcite within the grainstones and packstones (Fig. 5.24a-d). Whilst the wackestone has a micritic matrix (45-50%; Fig. 5.24e.f), with sparry calcite present only within vugs and pores. In terms of grain composition, the rocks are quite uniform, containing between 15-25% ooids (including superficial ooids), with subordinate, subequal proportions of bioclastic debris and pelloidal grains (Fig. 5.24a,c,d). Detrital quartz is also quite abundant within the transgressive sands, making up between 5-10% of the rock by composition (Fig. 5.24b-e; Appendix C4).

Some of the sections (e.g. HQTR1') contain abundant rip-up clasts, and although no clay clasts where sectioned I have noticed them in field studies. The granules of biooosparite within section HQTR1' (Fig. 5.24e,f) are similar in composition and preservation to sections of the underlying limestones (cf. Appendix C3; Figs. 4.8 & 4.9, Chapter 4) and clasts preserved within the clay-rich palcosols. It is likely that these intraclasts were derived by erosion of the underlying clay and carbonate regolith soils, during deposition of the carbonate sand.

The transgressive sands are largely replete of any early marine cements, such as radial-fibrous calcite or low magnesium syntaxial overgrowths. Most of the grainstones and packstones are cemented by a drusy ferroan sparite which preserves a low primary interparticle porosity (Fig. 5.24a-d). Secondary intercrystal and mouldic porosity is developed within some of the more weathered sections (HQTR1-1). The wackestone, HQTR1', is largely indurated by the micritic matrix and some opaques (Fig. 5.24e,f). It has a moderate amount of vug and fenestral porosity (i.e. birdseyes), which have been infilled by the slightly ferroan sparite cement. In contrast, the cementation history of the karstic intraclasts within this section are much more complex. Intact fibrous-radial and syntaxial overgrowths are seen upon grains and 45-50% sparite intergranular cement. The clasts are coated with a brown opaque-filled micritic weathering rind (Fig. 5.26e,f) following karstic and terrigenous chemical weathering (section 4.4).

Except for within the limestone intraclasts, the preservation of the cements within the limestones are fairly good, the ferroan sparite shows little evidence of chemical oxidation and the secondary porosity is consequently quite low. However, the grain preservation is quite variable ranging from extremely well preserved large low magnesium bioclasts (Fig. 5.24b) to almost totally micritised grains (Fig. 5.24a, c,d). Over 90% of all grains within the samples showed micritic envelopes and between 50-65% of them were partially micritised, usually with the development of a micritic weathering rind (Fig. 5.24a,c-f). Decalcification is quite low within these weathering rinds and granules of ferric oxyhydrite opaques are rare (<5%; Appendix C4). Ooids are calcitised and show well-preserved concentric zoning from inner calcitic to outer ferroan rich lamellae (Fig. 5.24c,d). Some also exhibit a well-preserved radial structure in cross-polarised light. About 2-5% of all bioclasts are calcified and the same amount have been wholly or partially dissolved, and infilled by sparite cement (Appendix C4).

5.13.2. Depositional environment of overlying sand unit

The carbonate sand unit is considered to have been deposited during inundation of the coastal marshes by returning marine conditions (Vaughan, 1988, 1989, Metcalf *et al.*, 1992). This probably occurred through a process of shoreface retreat landward, initiated by either tectonic subsidence or a slow rise in sea-level (section 6.3). The sands were probably deposited under intertidal conditions and can be considered to be a palaeobeach horizon. This high energy depositional environment is indicated by the well sorted and predominantly flat-laminated nature of the horizon, and the presence of reworked grains from the underlying soils and karst. The heavily micritised allochems also suggest active coastal erosion, within oxidised intertidal conditions. Following further drowning of the land and shoreface retreat these deposits were succeeded by the oolitic shoals of the overlying Chipping Norton Formation limestones.

A similar transgressive profile is seen in the modern coastline of northwest Florida (Hine *et al.*, 1988), where karstic limestones are overlain by thin coastal marsh soil horizons (rooted, organic-rich muds) which are abruptly coverd by a molluscan bearing, quartz-rich grainstone deposit. The coastal marsh and beach sands are 1-1.5m in total thickness and wave-attack or biocrosion removes this covering from over 50% of the coastal area following marine inundation, the sediments being preserved in dolines and depressions (Hine *et al.*, 1988).

Chapter Six Correlation and implications

6.1. Local correlation of the Hornsleasow deposits - a Chipping Norton Formation facies map

The Hornsleasow paleosols and palaeokarstic terrain provide a glimpse into an unique terrestrial ecosystem within the British Lower Bathonian. However, although the palaeokarstic horizon can be traced with some confidence along the face of the quarry, the clays are much less persistent and only occur in localised pockets developed within topographical lows. Clearly the lens described by Channon (1950) is laterally equivalent to the present unit, and further lenses may have gone unrecorded during earlier quarrying activity, as Huntsman's Quarries Ltd. workmen recall finding large bones and pockets of clay at this horizon on previous occasions. No recent (post-1988) quarrying activity has taken place at the site, and in the present economic climate, Huntsman's are unlikely to expand the workings in the near future. The site of the excavation is to be preserved for future research, but the karstic surface is rapidly degrading and the remaining clays have slumped into the hollow. If future research is to be carried out at the site, major re-excavation will be required.

The site is situated within a downfaulted block of Chipping Norton Limestone Formation, which is flanked on the north and south faulted sides by the older Inferior Oolite facies, and passes to the east into the underlying deposits of the Clypeus Grit (Fig. 2.3). No other active quarry sites occur within this northern-most outlier (Fig. 6.1) and therefore, close lateral correlation cannot be attempted. However, fault bounded outliers of Chipping Norton Formation occur within the northern Cotswolds to the south of the present site (Fig. 2.3), and it is to these outcrops where the search for contiguous terrestrial deposits was made. During the summer of 1992, extensive fieldwork was carried out in northern Gloucestershire and Oxfordshire in order to locate potential sites and attempt to correlate the terrestrial deposits recorded here, with the main Chipping Norton Formation outcrops further east. A review of the literature was also carried out, as many of the sites recorded in older reports are long disused, infilled or lost. The field localities I visited or mentioned in the following text are shown upon a locality and facies map of the northern Cotswolds (Fig. 6.1, modified after Sellwood & McKerrow, 1974).

6.1.1 Lateral continuity of the Hornsleasow paleosols and palaeokarstic land surface to the west of the Moreton Swell structure

In the immediate vicinity of Hornsleasow (Fig. 6.1) upon the western flanks of the fault bounded Moreton swell structure, the Chipping Norton Formation consists of a fairly

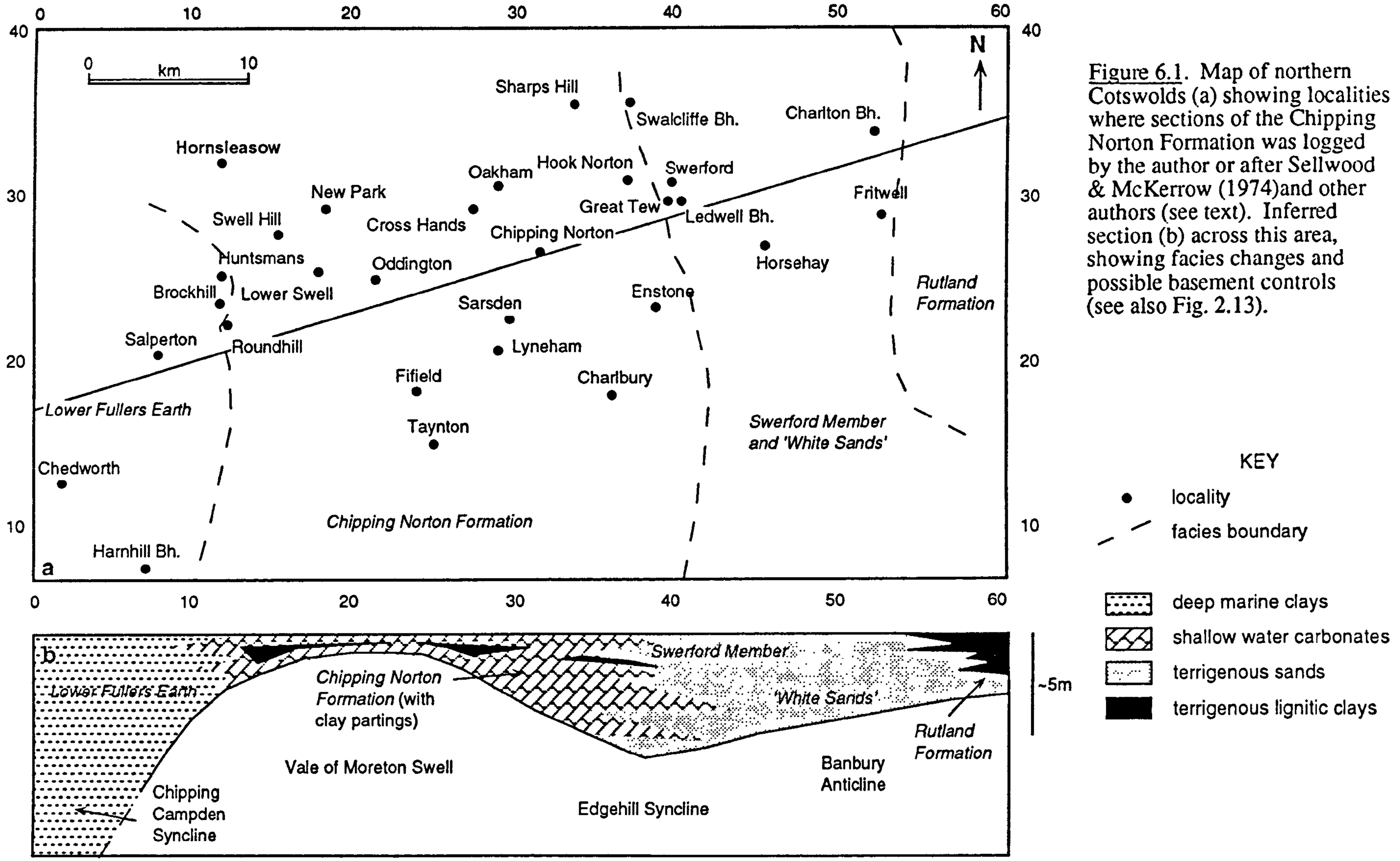


Figure 6.1. Map of northern Cotswolds (a) showing localities where sections of the Chipping Norton Formation was logged by the author or after Sellwood & McKerrow (1974)and other authors (see text). Inferred section (b) across this area, showing facies changes and possible basement controls (see also Fig. 2.13).

thick sequence of oolitic limestones and interbedded marls and clays. At Swell Hill Quarry, near Condicote (SP146270; Fig. 6.1) the oolites are quartz-rich, and the sedimentary textures observed there indicate rhythmical storm deposition in a shallow marine environment (Sellwood & McKerrow, 1974). However, no clay lenses were observed at this locality. The lower limestones (cf. Hook Norton Member) of the Chipping Norton Formation are more argillaceous than those above, and they contain much ground up lignite, pieces of coalified wood, clay drapes and pipes (Horton *et al*, 1987, S.J.M. pers. obs.). It is worth noting here that Richardson (1911) mentions clay beds at Lower Swell (SP174256, south of Hornsleasow), below the Chipping Norton Limestone.

Although many of the sections in this area have been infilled the New Park Quarry, Longborough (SP176282; Fig. 6.1) near Stow still shows about four metres of Chipping Norton Formation oosparite and bioclastic limestones (pers. obs.). The quarry is an SSSI for fossil reptiles (Benton & Spencer, 1995), yielding a fauna of dinosaurs and marine crocodiles in the period between the early 1900's and the late 1940's (e.g. Richardson, 1929, Gardiner, 1935, 1937, 1938, Reynolds, 1939). The section has been described by Richardson (1929) and following the dinosaur digs and further excavation in the 1930's, it was re-described by Arkell & Donovan (1952, p.249, their measurements have been converted to metres for clarity) as:

"Thickness

Chipping Norton Limestone

6. Flaggy oyster limestone. *Exogyra (=Praeexogyra)* sp.: 0.61m

5. Marl with oysters, chiefly *Exogyra* sp.: 3.04m

4. False bedded, white, shelly oolite: 5.49m

?Roundhill Clay (=Hornsleasow Clay)

3. Impersistent seam of clay and locally sandy marl with small irony claystone pellets: 0-0.20m

Hook Norton Limestone

2. Nodular buff limestone, with dark fossil bones 0.3-0.5m down. *Parkinsonia neuffensis* auct. from this bed: 1.52m

1. Brownish and buff limestone with sandy marl partings: 1.52m"

When I visited the site in 1992, the one remaining section in the northern face of the quarry showed up to three metres of the bioclastic flaggy bed 4 overlying a clear erosion surface upon the nodular oosparites of bed 2, and more massive oosparite (bed 1) at the base of the workings. The impersistent clay seam (bed 3) was not present in this section, although at the southern-most end and in the northwest part of the section thin (20-50mm) sandy brown-clay partings were present at this surface. The surface on which the

clay lies is undulose and has an iron-stained 'knobbly' appearance. I interpret the surface and overlying clays as equivalent to the palaeokarst and paleosols at Hornsleasow. This is further supported by the presence of black pebble-like clasts consisting of manganese, limonite, homiolithic oolite and phosphate, being reported from within the nodular limestone directly underlying the surface (bed 2; Richardson, 1929). The fossil bones have been reported from several horizons, Richardson (1929) reports that most of the early finds (crocodilians) were recovered from bed 4, and Reynolds (1939) implies that those collected in the 1930's came from the same bed. However, Arkell & Donovan (1952) clearly identify their bed 2 as the source of these specimens, which include terrestrial dinosaur remains (Benton & Spencer, 1995).

Further to the west, sections at Huntsman's (SP 1225), Brockhill (SP 133236) and Salperton Quarries (SP078211) (Fig. 6.1) show that the Chipping Norton Formation is entirely absent and the overlying shallow marine Charlbury Formation rests conformably upon the Lower Fullers Earth clay (Metcalf & Underwood, in prep.). This is also the case in the southern Cotswolds, with typical sections in the Chedworth railway cutting (SP052139; pers. obs) and the Harnhill (SP070005) and Meysey Hampton (SU1200) boreholes (R.J. Wyatt pers. comm. 1993) near Cirencester (Fig. 6.1) showing the Charlbury Formation overlying the deep marine clays of the Lower Fullers Earth (Metcalf & Underwood, in prep.).

6.1.2 Lateral continuity of the Hornsleasow paleosols and palaeokarstic land surface upon the Moreton Swell structure

The Chipping Norton Formation sequence is markedly reduced over the tectonically controlled Moreton swell to the east of Hornsleasow, making correlation across this structure almost impossible. For instance, the section reported by Richardson (1911) at Oddington (SP212256; Fig. 6.1) is about two metres, and southwards from here the sections of Chipping Norton limestones are never greater than three metres (Sellwood & McKerrow, 1974). The quarries at Taynton (SP241145) and Fifield (SP2418) have long been infilled, but Sellwood & McKerrow (1974) report that the Taynton quarry showed three metres of highly burrowed limestone with clay partings in the 1960's. They interpreted the sequence in this region as high energy oolitic banks similar to those forming in the Bahama Banks region at the present (Sellwood & McKerrow, 1974). Older sections from these quarries report that the Chipping Norton Formation limestones are sandy and contain blacks specks (?carbonised plant matter) (eg. Richardson, 1933), suggesting a nearby terrigenous influence.

6.1.3 Lateral continuity of the Hornsleasow paleosols and palaeokarstic land surface to the east of the Moreton Swell structure

The Chipping Norton Formation sediments are at their thickest east of the Moreton Swell in the northern Cotswolds (Fig. 6.1), where the formation reaches up to eight metres in places (Horton *et al.*, 1987), although there is still much evidence that the structure exerted some tectonic control upon sedimentation in this area. For instance, at Oakham (SP281307; Fig. 6.1), the Chipping Norton Formation section is capped by an oyster-encrusted hardground, which Sellwood & McKerrow (1974) consider as evidence for a reduced sedimentation rate in the northern Cotswolds on the eastern flanks of the swell structure. The eight metre section at Oakham rests unconformably upon the Inferior Oolite (Richardson, 1911) and the basal five metres consists of silicic biooosparites, which are very poorly fossiliferous, except for the presence of disseminated lignitic matter (Sellwood & McKerrow, 1974). The quartz sand and lignite content decreases toward the top of the section, but lignite-lined *Thalassinoides* burrows are reported within the higher cross-bedded units (Sellwood & McKerrow, 1974, p.199). These beds contain lenticles of shell-rich debris and Sellwood & McKerrow (1974) consider that they were deposited under a storm sedimentation regime in very shallow water, possibly with significant terrigenous input. In the working quarry at Cross Hands near Chastleton (SP270289; Fig. 6.1) the lower portions of the four metre section also show large plant fragments and disseminated lignite (Horton *et al.*, 1987; pers. obs., 1992).

Toward the eastern limit of the Moreton swell, Sellwood and McKerrow (1974, p. 198) described a clay deposit at Sarsden (SP 300266; Fig. 6.1), which Arkell (1947, p. 65) had erroneously assigned to the Sharps Hill Formation. This lignitic clay lens consists of a brownish green sandy clay, overlying a brownish grey clay which contains numerous oolitic clasts and is developed upon a *Lithophaga*-bored homiolithic limestone pebble bed (pers. obs. 1992), which Sellwood & McKerrow described as a hardground. The lens is described by Sellwood & McKerrow (1974, p.198) as within the "middle of the Chipping Norton Formation" and is situated around 1.5m above the base of the present section (pers. obs). Although the decalcified horizon underlying the lens could indeed by a hardground, as some of the pebbles also appear encrusted with small praeexogyrid oysters, it also could be correlated with the karstic surface at Hornsleasow, as the overlying Sarsden clay horizon contains the ?fresh-water gastropod *Bathonella*, '*Viviparus*' (Arkell, 1947, p. 65). It has been interpreted as a "temporary fresh-water pool within the carbonate-dominated area ... and attests to the extremely shallow-water nature of the carbonate region" (Sellwood and McKerrow, 1974, p. 198) in the region of the Moreton horst structure (Fig. 6.1). This clay horizon may also be laterally equivalent to the thin clay unit seen in a field to the northwest of Lyneham Barrow (2km southwest of Sarsden, SP 217211, Fig. 6.1) which has been described by Horton *et al.* (1987), and

which lies some 1.5m above the base of the Chipping Norton Formation. Horton *et al.* believed that the Lyneham clay could be correlated with the 'Roundhill Clay' (Richardson, 1929, cf. 'Roundhill Member', Sellwood & McKerrow, 1974) but this is unlikely, bearing in mind the abundance of impersistent clay horizons within the formation. The Roundhill Clay is by definition a thin (0.3m) band of oyster bearing clay occurring at the base of the formation and the type section is the Roundhill railway cutting (SP 125221; Fig. 6.1: 10km south of Hornsleasow and now largely overgrown). This unit is considered to thicken laterally into the Lower Fullers Earth clay facies in south Gloucestershire (Sellwood & McKerrow, 1974 and see below; Fig. 6.1).

6.1.4 Correlation across the 'Midlands shallows' region

Further to the east in Oxfordshire, the Moreton Swell exerted less of a control upon Chipping Norton Formation sedimentation. However, in this region considerable terrigenous input is supplied from the landmasses to the north and east as the Chipping Norton Formation grades laterally into sands, limestones and mudstones of the coastal marsh and lagoonal facies of the Rutland Formation (Fig. 6.1; Hallam, 1992). Sellwood and McKerrow (1974) described the facies variation of the lower Great Oolite (including the Chipping Norton Formation), and proposed a depositional environmental model for the succession from Oxfordshire and northern Gloucestershire. To the east (towards Northamptonshire), they describe a reduced-salinity, estuarine environment bordering the London-Brabant landmass.

The Chipping Norton Formation here (Swerford Member) is represented by cross-bedded silica sands with associated carbonaceous clay drapes, which interdigitate with sandy limestones and oolitic limestones of the upper beds of the Chipping Norton Formation (= Chipping Norton Member; Fig. 6.1). The Swerford Member sands were first described by Richardson (1911) and are present in sections around Chipping Norton, to the north and further east (Fig. 6.1). The old sections in the town reveal the upper beds of the formation in detail, and were recorded by Woodward (1894), and Horton *et al* (1987). Woodward (1894, p.324) recorded 3.66m of "brown and pale oolite, sharply jointed, current bedded, the lower beds tougher and siliceous" at Padley's Quarry in the southern area of the town (SP 318269). This quarry is famous for yielding cetiosaur remains in the 19th century, but is now almost completely overgrown and the basement beds are no longer visible (pers. obs., 1992). The type section for the formation at the council depot (formerly the Oxfordshire County Council and Workhouse Quarries, Sellwood & McKerrow, 1974; SP 318 274) is also now largely overgrown and the face is no longer stable, however Horton *et al* (1987, p.71) recorded over five metres of limestone, summarised below as:

	"Thickness (m)
Limestone; oolitic, shell-fragmental, with coarser bands of shell-fragmental limestone; well rounded quartz pebbles with a few limestone pebbles at three levels; thin false bedding:	1.55
Limestone; fine-grained oolitic:	1.02
Limestone; very fine-grained, thinly bedded with dendritic markings, darker in lower 0.36m:	0.56
Limestone (?Knotty Bed or *Trigonia signata* Bed); oolitic fine-grained, massive, with large trigonids:	0.91
Limestone; marly:	0.03
Limestone; fine-grained oolitic, with plant fragments in two massive beds:	1.22"

Clearly, both the plant fragments and the siliceous pebbles indicate a terrigenous input during deposition of these beds. Similar quartz-rich limestones occur in the Chipping Norton Formation sequence at Sharps Hill Quarry (SP 338358) and around Hook Norton (SP 355333) to the north of the town (Horton, 1977; Fig. 6.1).

To the south of Chipping Norton, in the disused quarries around the airfield at Enstone (SP 385250) and further south still at Charlbury Town Quarry (SP 370200) (Fig. 6.1), the sections are comprised of bioclastic and oolitic limestones with little terrigenous influence, except for abundant plant fragments. In this area the limestones represents deposition in relatively shallow-water marine conditions and are considered by Sellwood & McKerrow (1974) to be analogous to the modern day unstable sand-belts offshore from Florida and the Bahaman coastal region.

Further to the east in northern Oxfordshire, in the type region the sands of the Swerford Member are much more prominent and are there seen to "channel down deeply into the underlying Hook Norton Limestone" (Torrens, 1968, p. 230). Many of the early sections recorded by Richardson (1911) in describing the Swerford sands are now infilled or overgrown, even those described by Horton *et al* in 1987 (p.69) are no longer exposed. However, the Swerford railway cutting which pierces South Hill (SP358317) near Hook Norton (Fig. 6.1) still has a recognisable section, although somewhat obscured by vegetation and slippage. Here a lignite-rich brown clay lenticular horizon occurs within the section. The thin (10-30mm) lens overlies a decalcified oolite and siliceous sand deposit (1cm thick) which contains small bored and oyster encrusted oolitic pebbles (Horton *et al.*, 1987, pers. obs.). The surrounding limestone contains large (up to 0.15m) pieces of coalified wood, and appears speckled because of the abundance of lignite fragments (pers. obs.). The lignitic clay and bored pebble horizon

could be correlated with the Hornsleasow palaeokarstic surface and overlying soils, but they may also represent localised reduced sedimentation and hardground development within the marginal marine conditions.

The northern end of the tunnel has a more complete section and this has been fully recorded by Richardson (1911, p.213) and Horton *et al.* (1987, p.70), and is given here with minor alterations:

"Thickness (m)

Swerford Beds (=Swerford Member, Chipping Norton Formation)

Plant Bed; limestone, flaggy white oolitic:	0.61
Sand; weathering white:	1.07
Limestone; bored in places:	0.53
Sand; brown and yellow, with intercalated sandy limestone; contains *Praeexogyra* sp. and *Plagiostoma* sp.:	0.51
Limestone; sandy with incipient 'pot-lid' concretion structure at base:	0.70-0.91
Sand; yellow and brown; serpulids and *Praeexogyra* sp.:	0.03
Limestone; sandy; top surface of bed waterworn, pitted and covered with oysters:	0.56
Sand; coarse and gritty:	0.05

'*Trigonia signata*' Bed

'Old Man Limestone'; hard, brown, sandy, with water-worn surface covered with oysters, pebbly at the base:	0.38

Hook Norton Beds (=Hook Norton Member)

Sand; brown and yellow, with occasional 'knots':	0.05
Limestone; brown, shelly, with *Pteria inaequivalvis*:	0.60
Clay; dark with a brown layer:	0.18
Limestone; shelly:	0.13
Clay; arenaceous:	0.05
'Plant Bed' (of Walford, 1883); limestone; brown sandy, full of brown lignite fragments:	0.10
Clay:	0.06
Limestone; brown, shaley and marl:	0.61
Clypeus Grit; seen:	0.51"

Similar sections occurred at Great Tew (SP 394291; Fig. 6.1) and were recorded by Richardson (1911, p.218), although now only minor exposures occur in this region revealing 0.5m of fine-grained bioclastic limestone, siliceous and rich in plant debris and

carbon specks (e.g. at SP 395291), which are probably equivalent to the upper beds of the Swerford Member, recorded by Richardson as the 'Plant Beds'. An extremely nodular buff-coloured, fine-grained biooomicritic limestone occurs in the rubbish tip formerly known as the 'sand' pit between Little Tew and Ledwell (SP 395286). These beds of the Swerford Member consist of homiolithic limestone pebbles in close contact with one another, between which there are decalcified 'stylolitic' regions of dark reddish-brown, weathering dark yellow, iron-rich fine sand, giving the section a 'nodular' appearance. Neither plant debris nor any clay was observed at this site, but the nodular limestone is extremely similar to the decalcified region directly underlying the palaeokarstic surface at Hornsleasow. It is likely that the nodular limestone in the quarry was also formed through exposure, and subaerial or vadose weathering, but without further evidence it is not easy to elucidate whether the two horizons are equivalent. The Ledwell borehole about 10km east of Chipping Norton (SP 409282; Fig. 6.1) is the last place were the sandy bioclastic limestones and oolites of the Swerford Member are seen, further east the formation is replaced by the 'White Sands' of east Oxfordshire and Northampton (Fenton *et al*, 1994).

The 'White Sands' comprise pale grey, uncemented well sorted quartz sands (Horton, 1977) and are considered to have a terrestrial source. Much controversy has surrounded the lithostratigraphical position of this unit due to the absence of macroinvertebrate fossils. Previous authors have correlated the 'White Sands' with the Northampton Sands and Lower Estuarine Series (Bajocian) (e.g. Woodward, 1894; Thompson, 1924), with the Swerford Member (e.g. Arkell *et al.*, 1933; Sellwood & McKerrow, 1974) or stated that the sands represent two or more units ranging in age from the Lower Bajocian to the base of the Bathonian (e.g. Richardson, 1911). A good review of the stratigraphy of this unit is provided by Horton (1977), who provides lithological evidence in support of the first model. However, more recent palynostratigraphic investigations by the B.G.S. in this region have provided solid evidence that the 'White Sands' are equivalent to the Swerford Member (reported in Fenton *et al.*, 1994 and including sections in the Charlton (SP 529349) and Swalcliffe boreholes (SP 368359) and at Horsehay Quarry, Duns Tew (SP 455272) - all in Oxfordshire; Fig. 6.1.). Their palynological dating of a late Bajocian-Early Bathonian age for the unit also suggests that the lower part of the Chipping Norton Formation (i.e. the Hook Norton Member) also passes eastwards into the sands (Fig. 6.1). The diachronous nature of the 'White Sands' from Oxfordshire into Northamptonshire indicates younging toward the northwest (Fenton *et al.*, 1994), this factor is highly suggestive that the sediments were deposited in an embayment open to the north and southwest at that time.

The sands in the section at Horsehay Quarry and further to the east in the Fritwell railway cutting (SP 515289; Fig. 6.1.) are slightly felspathic, iron-stained and contain

abundant clay and lignitic partings (Sellwood & McKerrow, 1974). These coarsely clastic sands and coal seams have been described as "marginal marine" by Hallam (1992, p. 338), and are thought to pass eastwards into Stamford Member sands of the Rutland Formation (formerly the 'Upper Estuarine Series': Bradshaw, 1978; Hallam, 1992). These are interpreted as channel-fill deposits, which are possibly of deltaic or regressive marsh affinities fringing the low-lying London Brabant landmass. There is much land-derived plant debris, but the fauna within the sequence is restricted, indicating a low-salinity marginal-marine environment.

6.2. The source of the Hornsleasow clays - a volcanic or fluvial origin?

6.2.1. Carbonate rendzina and *terra rossa* soils

Although it is fairly clear that the clay paleosols and the underlying palaeokarst are intricately linked, the formation of such deep clay units over a fairly pure carbonate cannot be simply explained by simple karstic weathering and rendzina soil formation. Most rendzina soils are incredibly thin, and are composed of the insoluble residues from weathering of the underlying limestone. Weather-resistant residues of the limestones and allochthonous terrigenous material such as haematite, limonite, manganege and titanium oxides and aluminosilicates (kaolinite, illite and smectite) can accumulate upon such horizons and in tropical environments, the haematite imparts a characteristic red coloration leading to the term *terra rossa* soil formations (Davies, 1991). As limestones tend to be so pure, the allochthonous component tends to be the most important constituent of *terra-rossa* soils, and may be water- or air-borne (Foos, 1991). In modern tropical karstic terrains with little run-off, the wind-transported 'continental red-dust' which blows off the Sahara Desert is the common constituent of the *terra-rossa* soils which form (e.g. Foos, 1991). Airborne particles can also be introduced from aerial volcanic eruptions.

Terra-rossa soils tend to be very thin, mostly less than 0.2m in thickness, although some may reach up to a metre in deep karstic solution pits (Foos, 1991). They comprise a plant-litter 'O' horizon at the surface, a stony 'B' horizon, with aggregates of clay, oxides, limestone pebbles and rootlets and a 'Ck' horizon of pedogenically altered limestone (Foos, 1991). The sandy haematitic pebble horizon overlying the rubbly palaeokarst surface is considered to be such a soil and this layer is virtually clear of clay material. Carbonised and limonite-replaced plant material and the remains of rootlets at the surface of the karst confirm the presence of an 'O' horizon. However, the overlying clay-rich 'BC' or regolith horizon is also part of the soil, but the features and clay-mineralogy of this are intermingled with features of the overlying Grey Clay horizon. Similar ancient terra-rossa soils have been described from the Carboniferous (Riding and Wright, 1981) and the Jurassic of Portugal (Wright and Wilson, 1987).

In presenting analyses of the profiles of the two clay-soil horizons, the Grey and Green Clays, it became clear that these layers could not have arisen from dissolution of the underlying karstic limestones and airborne particulate introduction alone. Two possible sources were investigated for the Hornsleasow clay units, the first is a fluvial origin arising from the north or east and second is a volcaniclastic source material originating from the west and southwest. These two hypotheses are outlined below and summarised upon the block diagram, Figure 6.2.

6.2.2. A fluvial origin?

Upon investigation of the sand component of the clays, it was found that these contained up to 20% fine silica sand (section 5.8). This component can be further subdivided into two fractions. The first component consists of dull-surfaced 'metamorphic' grade quartz and felspathic grains. In thin section these characteristically show undulose extinction and some degree of weathering (Fig. 6.3a; section 5.8). This component is almost certainly derived from reworking of ancient metamorphic complexes and has a terrigenous origin. Of the siliceous sand component within the clays, the terrigenous sand makes up around 15%. The second, rare sand fraction is considered in more detail below (section 6.2.3). The silica-sand could have two sources, it could be derived from weathering of the underlying limestones and/or it could have been introduced into the hollow with the clay deposition.

To the east of the Moreton Swell structure both the lower and upper beds of the Chipping Norton Formation are quartz-rich sediments. The Hook Norton Member limestones characteristically contain drifted plant debris and siliceous clasts and the Chipping Norton Member limestones grade laterally into the terrestrial influenced Swerford Member in northeast Oxfordshire (Fig. 6.2). The sands of the Swerford Member have not been described in adequate petrological detail (cf. Sellwood & McKerrow, 1974) for a comparison to be made. Those of the contiguous 'White Sands' have been described and are said to contain sub-angular to sub-rounded clear quartz grains with very small proportions of chert, quartz with undulose extinction (i.e. metamorphic quartz) and feldspars (Horton, 1977). This is similar to the composition of the Hornsleasow sands, but Horton (1977) makes no mention of a shattered quartz component (section 6.2.3) and the scarcity of metamorphic quartz in the 'White Sands' is also different. He states (1977, p.152) that the 'White Sands' are orthoquartzites to sub-greywackes in petrographic composition, and that they also contain limonitic clasts, abundant ground up lignite and little carbonate. They are considered by Horton (1977) to represent accumulation in fluviatile and marsh environments (Fig. 6.2).

However, the quartz-sand content of the Chipping Norton Formation limestones decreases west of the Moreton-in-the-Marsh swell and at Hornsleasow it is only 10% within the limestones surrounding the lens. It is therefore extremely unlikely that the

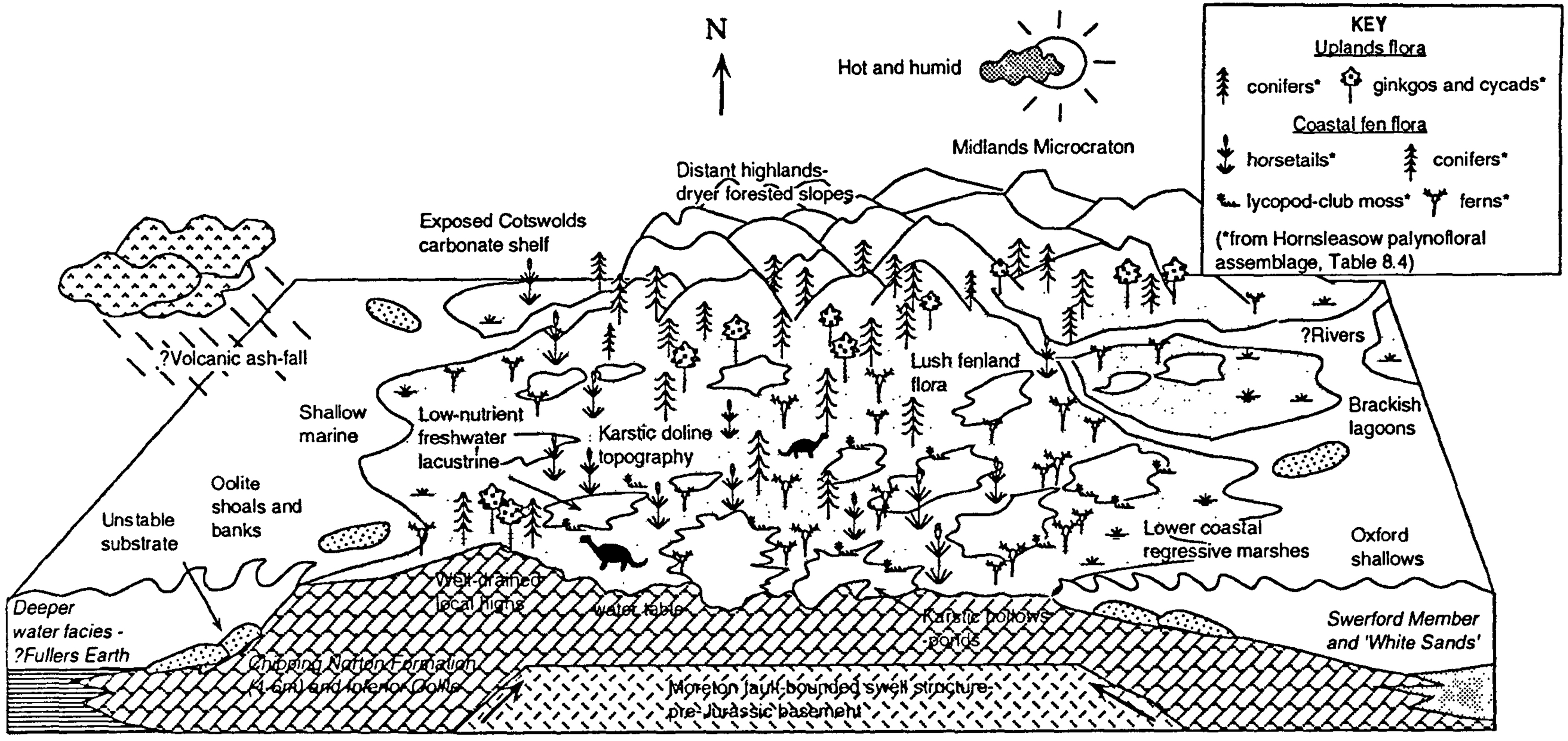


Figure 6.2. Proposed palaeoenvironmental and palaeoecological reconstruction of the Lower Bathonian (*zigzag* Zone) uplift and karstification of the Cotswolds-Weald carbonate shelf around the Moreton-in-Marsh swell structure. The climate was probably humid and fairly warm, and the depositional environment similar to the regressive marshes, brackish- to fresh-water lagoons and karstic ponds seen in the Florida Everglades at the present. North (N) is indicated.

Hornsleasow clays represent a westerly extension of the Swerford Member/'White Sands' facies. However, they could be derived from other terrigenous deposits. Although a fluviatile regime is not directly suggested for the sediments of the Hornsleasow hollow, the karstic region surrounding the doline may have had a drainage system which may have been subjected to periodic flooding events (Fig. 6.2). The palaeocurrent analyses of the hollow sediments (section 5.2) seemed to indicate a general palaeoflow direction orientated either east-west or north-northwest to south-southeast. This suggests that the Grey Clay at least, could have initially been deposited by flooding derived from either the northern Midlands cratonic area or from the Oxfordshire shallows (Fig. 6.2). If the former hypothesis is correct this would suggest that not only was the area around Hornsleasow emergent at this time, but that this region was connected to the northern landmass. However, if the latter is true this would indicate possible reworking and deposition of 'White Sands' sediments over the Moreton Swell. As this structure seemed to have exerted some control upon Chipping Norton Formation sedimentation and may be implicated in the emergence of the limestone around Hornsleasow, this would seem unlikely.

6.2.3 The Lower Fullers Earth Clay Formation as a possible parent material

The most obvious anomaly between a carbonate origin for the paleosol units is their almost exclusive clay mineralogy. Analysis by X-ray diffraction (XRD) procedures (Appendix D) suggest that the Grey Clay is composed of a matrix of illite and kaolinite (Metcalf *et al.*, 1992). An earlier analysis of this unit also revealed the swelling layered clay, smectite (Vaughan, 1989), but this statement was not supported by a XRD trace and the original analysis was written in an undergraduate report (R. Jones, pers. comm., 1991) which has since been lost! There is even some discussion that the undergraduate may have got the samples mixed up (D. Dartnall, pers. comm., 1993) and there is substantial need for a further analysis to be carried out in the future. The second analysis shows the XRD trace for the overlying Green Clay, which is essentially composed of a clay matrix of illite, with subordinate amounts of kaolinite. This was originally described by Vaughan (1989) as an illite/chlorite clay, based upon the evidence provided by the undergraduate. Illite and kaolinite clays are predominantly derived from terrestrial weathering of minerals, for example the Lias clays of Dorset are composed of a mixture of these two mineralogy's (Hallam & Sellwood, 1968) and the Jurassic illite/chlorite/kaolinite clay soils formed on top of karstic limestones in Portugal are also detrital in origin (Riding and Wright, 1981). However, illite/smectite mixed layering is common within clay-rich soils believed to have originated from the weathering and early pedogenic alteration of volcanic ash. Similar horizons are described in the Dinantian palaeokarstic terrain's of north Wales and Derbyshire (Davies, 1991, Walkden, 1972, Robinson & Wright, 1987) and are termed 'bentonites' (Walkden, 1972).

The second feature within the clay paleosols which indicates derivation from an external, non-terrigenous source material, is the abundance within the soils, of shattered quartz grains (Fig. 6.3b). This component make up around 5% of the sand fraction and are typically sub-angular to angular grains with a vitreous lustre. In thin section, the grains can be seen to be anhedral (Fig. 6.3b) to sub-euhedral lathic (Fig. 6.3c) crystals of predominantly quartz, although a felspathic (probably potash feldspar) component is also present. The grains show a fine mosaic of cracks (Fig. 6.3b,c) and exhibit uniform extinction in cross-polarised light. They are largely unweathered, unlike the terrigenous or metamorphic siliceous grains (Fig. 6.3a). Such crystals are found within volcanic tuffs formed in Surtseyan-type eruptions, were the pyroclastic material has been ejected through a crater-filling lake or glacier (S.J. Matthews, pers. comm., 1994). The passage through a body of water or ice, causes rapid quenching of the melt and any early-formed crystals within the magma are shattered (S.J. Matthews, pers. comm., 1994). Figure 6.3d shows a crystal of plagioclase feldspar within a Surtseyan tuff erupted from the active ice-filled Lascar volcano, northern Chile (section 'LAS 199' supplied by S.J. Matthews, 1994). The shattered crystals within the Hornsleasow clays could only have an explosive volcaniclastic origin.

The most obvious choice for a possible parent ash-fall deposit, would be the Lower Fullers Earth Clay Formation of the northern Cotswolds. This unit is part of the extensive Bathonian Fullers Earth 'Group' of southern England (Cope *et al.*, 1980a). Although some earlier authors (e.g. Newton, 1937) considered the blue-grey clay horizons of the Fullers Earth Clays to be sedimentary marine formations, it has now been unequivocally established that they are the weathered remains of subaqueous ash-fall deposits or bentonites (Hallam & Sellwood, 1968, Jeans *et al.*, 1977). The Fullers Earth sequence consists of an upper and lower clay horizon separated by a marly limestone in Dorset, Somerset and Avon and two limestones and an intervening clay band in the southern Cotswolds (Arkell & Donovan, 1950). In the northern Cotswolds, and significantly around Burford and Stow (less than 8km from Hornsleasow), the lowest clay unit lenses out and passes into the Chipping Norton Limestone Formation (Arkell & Donovan, 1950). The limestone comes in beneath the diminishing Lower Fullers Earth Clay Formation around Burford and Stow, and thickens as the clay unit diminishes to the north and east (Figs. 3.2 & 6.1; Richardson, 1911, 1929).

The petrology and mineralogy of the Lowers Fullers Earth clays has been determined by Hallam & Sellwood (1968) and Jeans *et al.* (1977). The clays are composed of layers of swelling montmorillonite and mixed illite/montmorillonite. Minor constituents of the clays are shards of unaltered devitrified glass, pumice, cracked feldspars, and quartz. Land-derived detritus and bioclastic material also occur within some sections of the Fullers Earth and the clays are well bioturbated. The clay mineralogy and bulk chemistry indicates subaqueous alteration of a trachytic ash in the

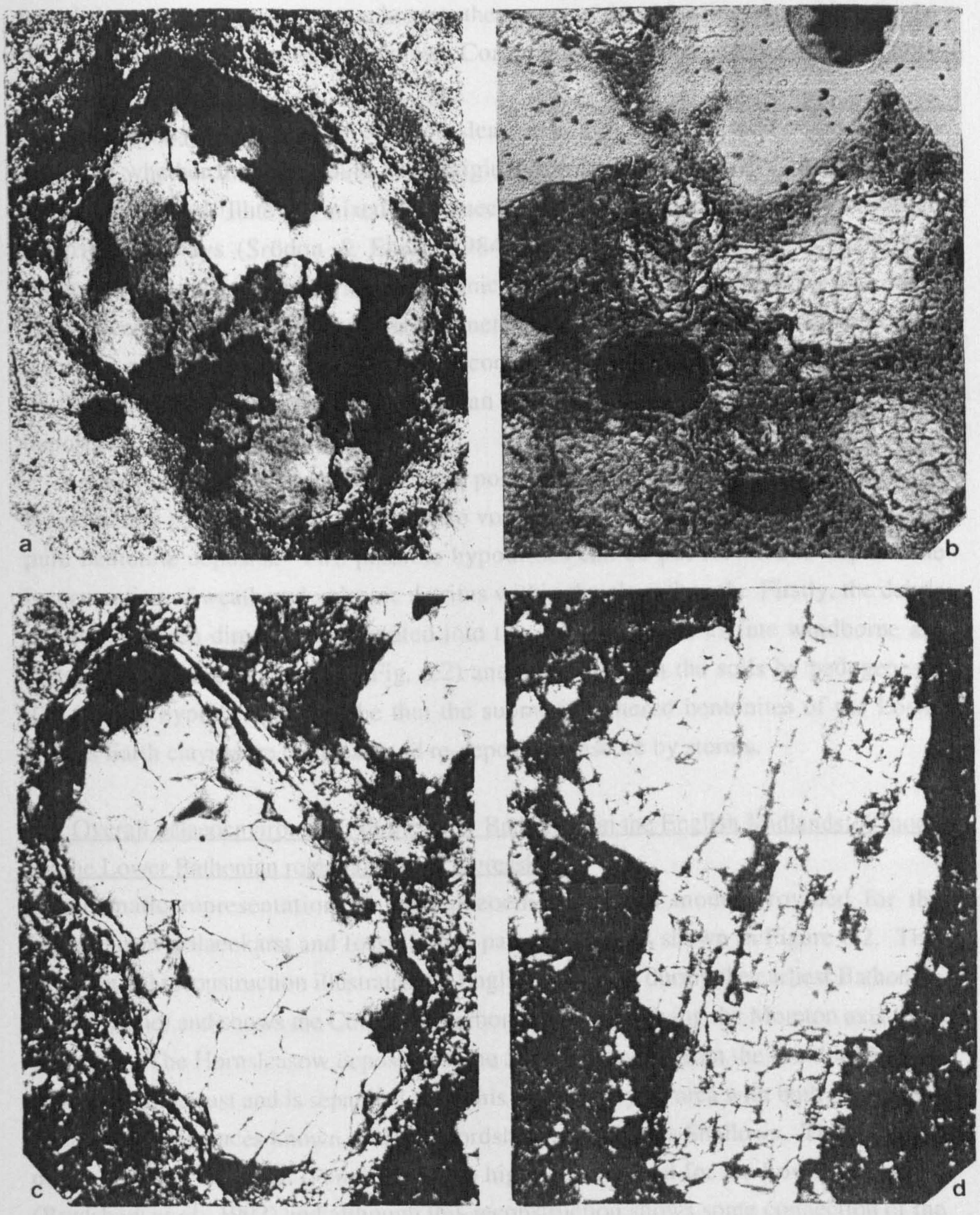


Figure 6.3. Photomicrographs of petrographic thin sections of (a) terrigenous quartz showing undulose extinction; section HQ S10-3, cross-polarised light; (b) shattered volcanic quartz; section HQ S10-1, plane-polarised light; (c) shattered volcanic quartz with uniform extinction; section HQ S10-3, cross-polarised light; (d) shattered volcanic quartz with uniform extinction; section LAS-199, Surtseyan-type eruption, Lascar volcano, Chile, cross-polarised light. Field of view. All sections were photographed under blue-filtered light.

reducing and alkaline environment of the marine basin (Jeans *et al.*, 1977). Although the actual site of the volcano(es) are unknown, the source of the volcanics is considered to be off the coast of the southern landmass, Cornubia (Cope *et al.*, 1992) and probably submarine.

The clay mineralogy of the Hornsleasow palcosols is not well constrained to elucidate whether the soils could have originated from the same source as the Lower Fullers Earth clays. Illite and mixed illite/smectite or illite/kaolinite clays can be formed in a variety of ways (Srödon & Eberl, 1984; Romano & Spears, 1991), including submarine and subsoil weathering of volcanic ash. However, illitisation may also occur during deep hydrothermal diagenesis and metamorphism. Romano & Spears (1991) do provide a general rule that if the sediments contain a mixed mineral assemblage then this implies a source inhomogeneity, rather than formation from weathering of a pure ash deposit.

Taking into consideration this last point, it would seem clear that although the Hornsleasow paleosols may contain some volcaniclastic debris they are unlikely to be pure bentonite deposits. Two possible hypotheses can be put forward to explain the incorporation of weathered volcanic detritus within the clays though. Firstly, the debris could have been directly incorporated into the clay sediments as fine windborne ash deposits following an eruption (Fig. 6.2) and altered within the soils by pedogenesis. The second hypothesis would be that the submarine altered bentonites of the Lower Fullers Earth clays were reworked and re-deposited onshore by storms.

6.3. Overall palaeoenvironment of the early Bathonian in the English Midlands: A model for the Lower Bathonian regional marine regression

A schematic representation of the palaeoenvironmental model provided for the Hornsleasow palaeokarst and fossiliferous palcosol units is shown in Figure 6.2. This hypothetical reconstruction illustrates the English Midlands during the earliest Bathonian (*zigzag* Zone) and shows the Cotswolds carbonate shelf overlying the Moreton axis to be emergent. The Hornsleasow deposit is some 50-60 km away from the London-Brabant landmass in the east and is separated from this land area by an area with both marine and fresh-water influences known as the Oxfordshire or Midlands Shallows. The shoreline of the northern landmass (Pennine) is also highly conjectural for the Lower Bathonian (Bradshaw *et al.*, 1992) and although this reconstruction shows some connection of the exposed carbonate shelf with the proto-Pennine mountains, this is not based on actual evidence. However, some connection with a more substantial land area is indicated by the megafauna discovered at the site (Chapter 9). The palaeofloral reconstruction and the palaeoclimate is based upon the evidence supplied in Chapters 8-9.

6.3.1. Tectonic controls versus sea-level controls

The Bathonian stage in the British Isles is a relatively regressive phase in the primarily marine strata of the Jurassic. Sea-level curves suggest that there was a stillstand in the early Bathonian (section 1.2.6). During the Bathonian stage, southern Britain was covered in the shallow seas of a stable marginal platform basin (Bradshaw *et al.*, 1992). The Chipping Norton Formation was deposited upon the wide, expansive Cotswold-Weald shelf (Fig. 3.2). However, the localised extent of the terrestrial conditions developed at Hornsleasow, would suggest a local, probably tectonic control upon the retreat of marine conditions off the carbonate shelf in this region, rather than a world-wide eustatic event.

Such a tectonic control can be explained by fault-block activity upon the Cotswolds-Weald shelf which is known to have been tectonically active during Lias and early Mid Jurassic times (e.g. Chidlaw, 1987; Mudge, 1978; Baker, 1981) (Fig. 3.2). The Moreton-in-Marsh swell is thought to have been a land area in the early Jurassic (Chidlaw, 1987) and this became an island again in early Bathonian times (Fig. 6.3). During the remainder of Bathonian times the Moreton-in-Marsh swell seems to have provided a waning influence upon sedimentation in the Cotswolds and by late Bathonian times the tectonic control in the region had ceased (Chidlaw & Campbell, 1988). At the same time there was a eustatic rise in sea-level and the Lower Cornbrash was deposited over the Great Oolite Group.

6.3.2. Palaeoclimatic implications of the Hornsleasow palaeokarst and paleosol units

Karst land forms are ubiquitous over many different climatic zones, and therefore, strict climatic interpretation of karst topography is not advised (Trudgill, 1985). Also the size of the outcrop at Hornsleasow cannot reveal the true picture of the whole karstic landform in the northern Cotswolds. Hence, any interpretation is based upon supposition and extrapolation of the features seen in the excavation site (e.g. Fig. 6.2). Jennings (1985) has stated that in general the overall morphology of karstic landscape is related to climate, for example, dolines and other karren forms tend to be more abundant, larger and well-defined in tropical regions. Whereas in cooler climates, karst topography is more subdued and there are fewer dolines (Retallack, 1990). Case-hardening cements, subsoil weathering and leaching of host limestones is mainly found in tropical humid areas (Wright, 1981). However, karst morphology can also depend on other factors such as cementation patterns in the host limestone and water-flow regimes, and the complexity of modern day karst terrains reflect this, for example the coastal karsts of south-west Florida (Hine *et al.*, 1988).

However, the features seen in the Hornsleasow palaeokarst horizon would tend to support other evidence (e.g. palaeobotanical evidence, section 8.3.4) that the climate in

the lower Bathonian was warm, tropical and rather wet (Fig. 6.2). Subsurface caverns are not present in the karstic sequence and this may be because of the overlying clay-rich soils. In tropical and seasonally wet regimes, such soils would retard the rate at which rain-water could percolate through to the underlying limestone bedrock and limit the effects of vadose dissolution to along the small stylolitic fractures (section 4.4.5). Retention of the water within the soils is exemplified by the gleyed Grey Clay paleosol at Hornsleasow and would have encouraged dissolution along the rock/soil interface, resulting in the decalcified and brecciated surface of the palaeokarst and gradational 'regolith' horizon. A similar sequence is seen in the Pleistocene karsts of Barbados, which underlie smectitic-rich palcosols (Harrison, 1977).

The hollow-filling fossil soils can also provide useful palaeoclimatic information, although it is true to say that soils often support their own microclimates, and this is particularly so for poorly-drained waterlogged soils, where oxygenation and temperature may be controlled by groundwater flow rather than atmospheric conditions. Nevertheless, good palaeoclimatic indicators, such as glaebules, soil textures, root morphology and clay mineralogy, are preserved in the Hornsleasow paleosols.

Firstly, seasonal palaeoclimatic conditions seem to have been prevalent during the formation of (at least) the Green Clay palcosol profile. Features such as the ferruginous rhizocretions, a mottled 'Ae' horizon, and calcareous nodules, and the presence of blocky ped structures and slickenslides in the Green Clay are all supportive of a pronounced seasonality. Some of these features also occur within the underlying Grey Clay, although this was waterlogged for most of the time of its formation. Therefore, the effects of seasonality would have been subdued by the soils' microclimate during maturation, and there is also the possibility of polygenetic overprinting during the formation of the comparatively well-drained Green Clay.

Most chemical weathering reactions which occur in soil profiles are controlled by temperature, rainfall and maturation time. Tropical soils are much more likely to be more deeply weathered than temperate or polar soils (Retallack, 1990). The general rule is that the more reddened and more clayey a soil matrix, the more humid the environment. The Hornsleasow soils are certainly clay-rich, and contain abundant oxidised iron, but are not reddened, because of their gleyed nature (Grey Clay) or from sea-water alteration (Green Clay). Clay minerals are weathering products of hydrolysis, and their mineralogy can be directly related to the availibility of moisture to the soil. Clay minerals occurring in wetter climates, where there is intensive humid weathering, are base-poor, and include kaolinite and halloysite (Singer, 1980). Whilst smectites, illites and chlorites occur in much dryer climes or those with a pronounced dry season (Singer, 1980). However, the clays are not found *in situ* in the hollow, and and thus, to identify a specific weathering environment from clay minerals not found in their place of genesis is extremely hazardous (Singer, 1980). A similar study was undertaken by Hurst (1985) upon the

clay paleosols of the Brora Coal Formation (Aalenian-early Callovian) of north-east Scotland. Previous authors (e.g. Hallam, 1973) had suggested a humid, possibly tropical climate, based upon the abundance of kaolinite in the soils. However, Hurst (1985) found that the clay mineralogies, kaolinite and mixed illite/smectite layering, were derived from contemporary reworking of local Devonian and Carboniferous sediments, and had no bearing upon the palaeoclimate of the Middle Jurassic. He suggested that the original clay mineralogies were stable in the mildly alkaline, non-intensive weathering at that time, and supported a subtropical, seasonal palaeoclimate, with variable humidity for the Middle Jurassic of north-east Scotland (cf. Hudson, 1980).

It is clear then that much more work needs to be carried out upon the mineralogy and provenance of the Hornsleasow paleosols, before a palaeoclimatic model can be fitted. However, the paleosols were not devoid of life, and it was therefore possible to glean much more information about their palaeoenvironment from the fossil remains contained therein (Chapters 7-10).

Chapter Seven Palaeontology - an introduction

7.1. Why study the Hornsleasow fauna

The palaeontology of the Hornsleasow paleosol units was studied after the sieving project at Gloucester Museum and Bristol University was nearing completion. In the five years of production, over 25000 fossils were recovered and curated into Gloucester City Museums collections, of which I examined over 2000 remains. Of this assemblage, 85-90% fossils were recovered from the Grey Clay paleosol and the rest from the Green Clay. The rough percentage abundances of vertebrate and non-vertebrate fractions in these contexts are shown in Fig. 7.1, but these figures are variable over the site, as for example, the vertebrate fraction can be lower than 5% in some areas and as high as 70% in others. In all, fifteen tons of sediment was sieved during the course of this study.

The fauna is extremely important for two reasons. The first is the relative rarity of Middle Jurassic terrestrial fossils. In terms of the terrestrial vertebrate fossil record and their evolution, the number of remains produced from the site makes Hornsleasow one of the most important British fossil localities to be discovered this century. In the succeeding chapters (Chapters 8 and 9) the fossil remains are listed and described in detail. Chapter 9 stresses the importance of the content of the Hornsleasow vertebrate fauna, and provides a synopsis of a paper to be submitted this year on the assemblage (Benton *et al.*, 1995). Chapter 8 is a summary of the non-vertebrate assemblage.

The second reason for researching the Hornsleasow accumulation, is its unusual preservation style and faunal content (Table 7.1). Paleosols are not usually known for their abundance of fossilised remains (Bown, 1979, Bown & Kraus, 1981 and references therein; Maas, 1984; Schultz *et al.*, 1955), although this may be a function of the failure to recognise such deposits as soil horizons in the geological record. The preservation of so much material within the Hornsleasow soils and the way it is preserved, can provide extremely important palaeoenvironmental information for the two paleosol units. For example, Fig. 7.2 illustrates the Eh/pH stability fields for fossilised remains within various natural freshwater systems, including soils. Both Hornsleasow paleosols had near neutral to slightly alkaline conditions (Fig. 7.2), as they contain fairly well preserved calcareous matter, including invertebrate fossils (Chapter 8) and phosphatic material in the form of vertebrate remains (Chapters 9-10) and coprolites (Chapter 10). However, the bulk of the well-preserved carbonaceous plant matter is found in the lower Grey Clay (Chapter 8), indicating that the soil was permanently waterlogged (Fig. 7.2). Whilst plant material in the Green Clay has been oxidised to iron

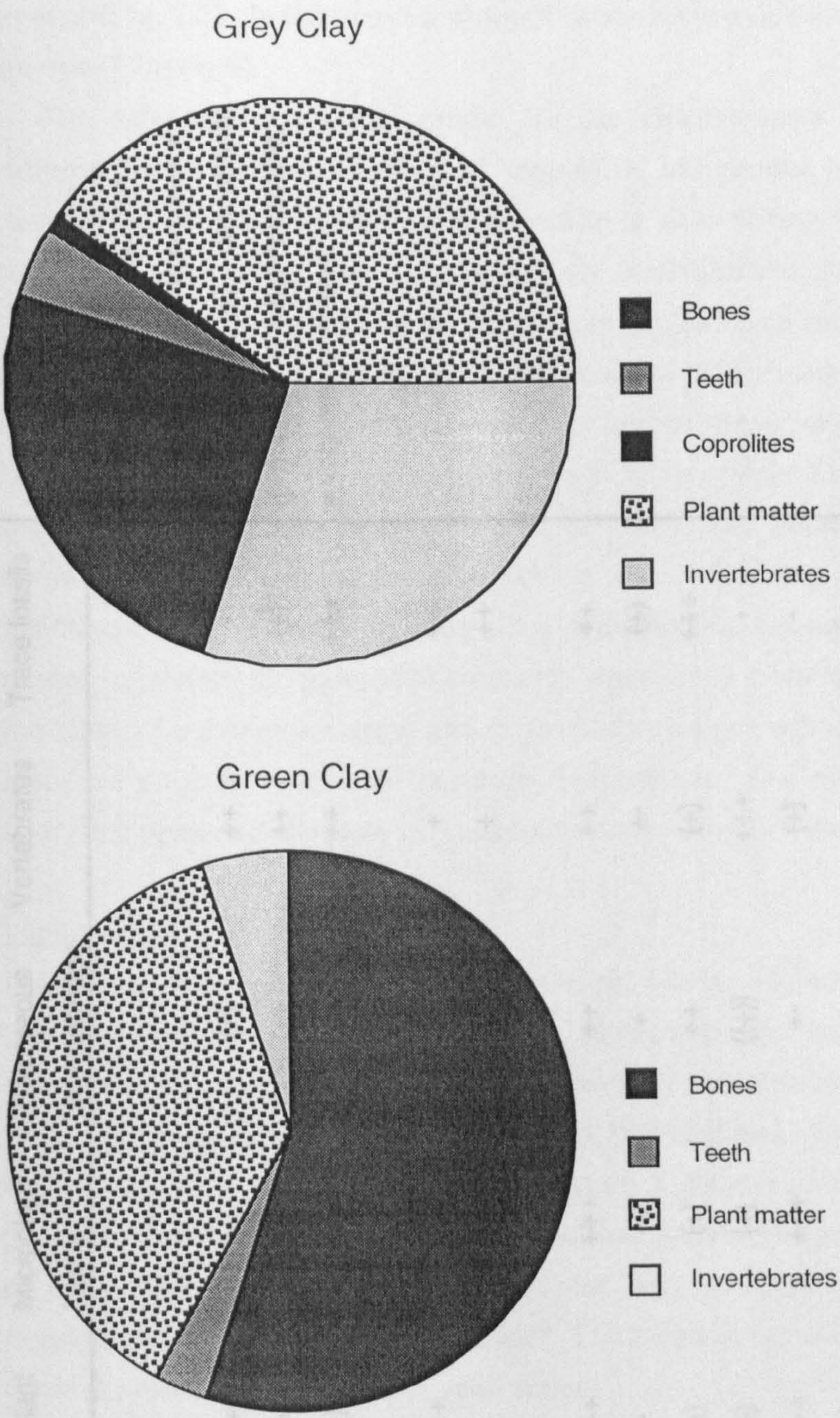


Figure 7.1. Rough abundances of fossil material for the sieved fractions recovered from the Green and Grey Clay paleosol units.

Palaeoenvironmental Context	Occurrences of terrestrial fossils				
	Macroplant fossils	Microplant fossils	Calcareous invertebrates	Vertebrates	Trace fossils
Lacustrine (small lake)					
Low oxygen					
deep	++	++	(+)	++	-
shallow	++	++	(+)	++	(+)
Oxygenated	(+)	++	+++	++	+++
Floodplain					
Poorly drained	+++	+++	+	+	++
Well drained	+	(+)	(+)	+	++
Fluvial/deltaic					
Abandoned channel	+++	+++	++	++	++
Crevasse splay	+	++	+	+-	(+)
Levee	(+)	(+)	++	(+)	+++
Karst	(+)	(+)	((+))	+++	-
Peat deposits	+++	+++	+-	(+)	-
Faeces and regurgitaes	(+)	+	+	++	-
Volcanigenic lacustrine	++	+++	++	++	+++

Relative abundances : + = present; ++ = common; +++ = very common; (+) = uncommon; ((+)) = rare; +- = highly variable; - = very rare or absent.

Table 7.1. Summary of the preservational modes of terrestrial fossils in certain terrestrial palaeoenvironments after Behrensmeyer & Hook (1992).

oxides and hydroxide compounds (Chapter 8), suggesting that the soil was fairly well oxygenated (Fig. 7.2). Indigenous plant fossils can also provide excellent palaeoclimatic information (Chapter 8).

The palaeoenvironmental model for the Hornsleasow paleosols represents deposition within a marshy karstic pool supporting indigenous aquatic, and terrestrial animals derived from the floodplain, followed by a slow silting and desiccation of the pool in Green Clay times and overprinting by well-drained soil forming processes (Chapter 5). Neither of these palaeoenvironments precludes an indigenous biota, but the shear abundance and diversity of fossilised remains within the excavated hollow at Hornsleasow suggests an unique palaeoecology and/or preservational environment for the fossils (Table 7.1). In the next sections I shall briefly explain the procedures involved and problems faced in trying to elucidate the palaeontology of a fossil accumulation. For a fossil assemblage is not the same as a living community of animals and plants, and hence, before providing clues to the ancient ecology and interplay of the representatives within the palaeoenvironment, based upon a study of the accumulation, one must firstly be able to recognise and remove any changes which have occurred to that accumulation after the deaths of the living individuals. The study of the modifying effects of time upon an individual or palaeocommunity is called 'taphonomy'.

7.2. Taphonomy

The word taphonomy is derived from the Greek words, *taphos* meaning burial, and *nomus*, meaning law (Efremov, 1940). However, the science of taphonomy encompasses much more than just the laws of burial for an individual fossil. It is widely regarded as the study of the variety of biological, physical and chemical processes which determine how and why a once-living form has been preserved as a fossil (Allison and Briggs, 1990) and encompasses a range of multidisciplinary techniques. In grass-roots terms, taphonomy seeks to answer the question: why was this fossil preserved, rather than being destroyed by the ravages of time? The original definition of taphonomy was given by I.A. Efremov (1940, p.85), who stated:

> "The chief problem of this branch of science [taphonomy] is the study of the transition (in all its details) of animal [and plant] remains from the biosphere to the lithosphere [geological record]".

The importance of studying the preservational styles and faunal content of a fossil assemblage is to elucidate whether the individuals preserved within that accumulation represent the individuals of a real palaeocommunity. In studying a fossil assemblage one is rarely sampling the original living community, but one that has been modified by post-mortem effects to the individuals of that community, or in the case of a temporally or spatially mixed assemblage several sampled communities (Table 7.2 & 7.3). This is especially true for the study of vertebrate preservation which is complicated by the fact

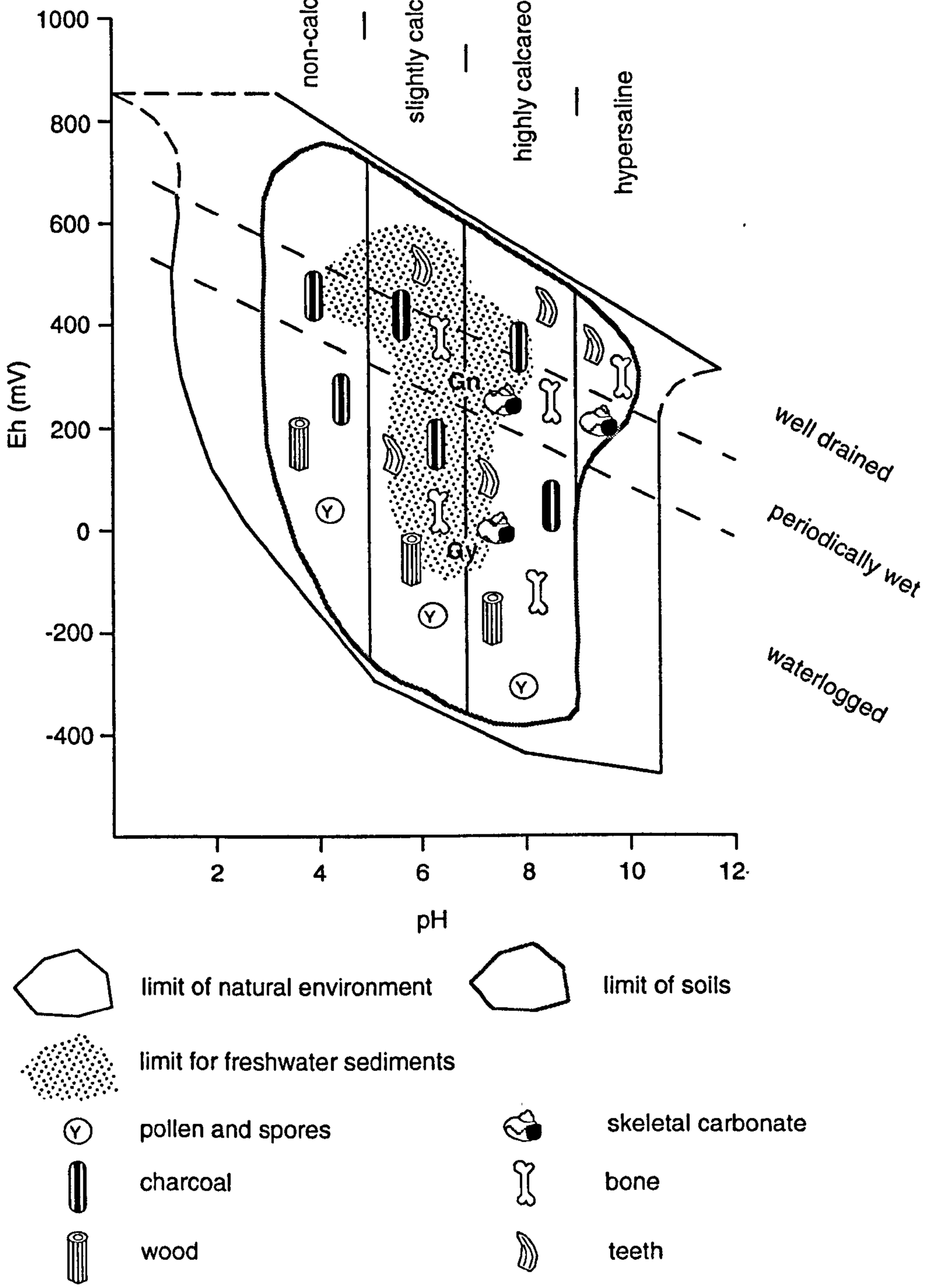


Figure 7.2. Stability fields in natural freshwater systems, including soils, for various fossil remains (after Retallack, 1990). The limits for the Grey (Gy) and Green (Gn) Clay paleosols are shown.

that each individual animal has over 150 bones, as well as scales and teeth; and in the case of reptiles, which continually replace their teeth throughout life, many hundreds of shed crowns produced before the death of the individual. In all but very exceptional preservation, these elements become separated after death of the animal, and undergo separate modification. Hence, in a mixed, fragmentary fossil assemblage every specimen must be assessed with equal care in determining derivation of the material and the preservational history of the assemblage (Metcalf, 1993). This means that a fragment of bone or shell is as important as an identified limb element or invertebrate test for palaeoecological reconstruction.

Taphonomy essentially serves to recognise the range of processes that change a once living entity or community into a research specimens or collection, to provide a full taphonomic history or profile (Table 7.2). In doing so, taphonomy provides the detailed knowledge of events which have biased a fossil accumulation, and therefore, stripping away this taphonomic overprint from the fossil record will enable an accurate picture of the biotic community to be made. Therefore, taphonomic studies should always precede palaeoecological reconstructions (Shipman, 1981).

However, although, taphonomy arises from a desire to elucidate the loss of information to the fossil record, it can also provide a much more positive input in the study of fossil accumulations. Most recent taphonomic studies have focused upon the positive aspects. For example recognising the type and reason for a taphonomic bias in the fossil record is much more positive than simply stating that the record is incomplete. Other positive aspects of taphonomic studies include recognising traces of predator/prey interaction (e.g. bite marks), sedimentary and depositional environments (e.g. abrasion, orientation of fossils) and fossilisation processes (e.g. geochemistry of enclosing matrix porewaters). In short, the study of a taphonomic overprint can provide the best explanation for the processes responsible for a particular fossil or assemblage being preserved (Gifford, 1981). Behrensmeyer and Kidwell (1985, p.105) reviewed the contribution of taphonomic research to palaeobiology and sedimentology as a whole, and gave an alternative definition of taphonomy as:

> "The study of processes of preservation and how they effect information [in the fossil record]."

7.2.1. History of previous research

The foundations of modern taphonomic research were laid down with those of palaeontology (Cadée, 1990). Most palaeontologists in describing fossil organisms, would make observations and theories on how the organisms might have become fossils. These early workers focused upon interpreting fossil deposits in terms of post-mortem processes that they were able to observe in the natural environment. Much of the best studies were performed in Germany between the two World Wars, and included the

monographic works on the studies of death, decay, disarticulation and burial of vertebrate carcasses by Weigelt (1929) and Richter (1928). It was Richter (1928) who coined the term *'aktüo-palaeontologie'* for the study of post-mortem processes on modern organisms and their relevance to the form of fossil assemblages. This concept of the present being the key to the past is known as *'uniformitarianism'* and is still regarded as one of the key parameters in taphonomic research (see discussion in Lyman, 1994).

However, taphonomy was not recognised as a separate study until the mid and late 20th century, after the defining work of Efremov (1940). Efremov recognised that diagenetic factors also bore relevance to fossil accumulations as well as biostratinomic and necrolytic effects. Much of the work following this in the 1950-1970's was directed towards analysing the effects of taphonomic bias or losses in the fossil record and how to eliminate it in palaeoecological reconstructions. This body of papers included works on marine invertebrates, such as those by Lawrence (1968) and Johnson (1957, 1960, 1962) and terrestrial vertebrates, such as those by Olson (1952, 1958), Shotwell (1955) and Voorhies (1969, 1970).

Actualistic studies became fashionable in the late 1960-1970's especially with the onset of the ideas that taphonomy could have a potential contribution to archaeological research (Gifford, 1981), and included work upon transportation (Toots, 1965; Voorhies, 1969), disarticulation (Hill, 1978, 1979a,b; Hill and Behrensmeyer, 1984, 1985), and weathering (Hill, 1976; Behrensmeyer, 1975, 1978) of modern animal skeletons (note that these studies mostly refer to mammalian vertebrate remains). The 1950-1970's also saw the introduction of statistical techniques being applied to taphonomic investigations, including the estimates of original numbers of living organisms by Shotwell (1955), Grayson (1978) and Holtzman (1979), and the modelling of large-scale biases in the fossil record by Raup (1976, 1979) and Sepkoski (1976).

For a more comprehensive review of the early literature of taphonomic research, the reader is directed toward Behrensmeyer and Kidwell (1985), Thomas (1986), Cadée (1990), papers in Donovan (1990), Briggs and Crowther (1990) and Allison and Briggs (1991), and specifically for vertebrate taphonomy, Shipman (1981), Behrensmeyer and Hill (1980) and Lyman (1994).

7.2.2. Current status and problems in taphonomy

In the past fifteen to twenty years taphonomic study has come along in leaps and bounds, in most cases palaeontological studies are not considered to be complete without some mention of preservational styles and possible taphonomic histories. Indeed it was in the mid-1980's when the concept of fossils possessing a taphonomic history was first defined by Andrews and Cook (1985), although proto-taphonomic histories were illustrated by Fagerstrom (1964) and Seilacher (1977). Taphonomic histories are also now understood as being complex and cumulative, i.e. the effects of a late taphonomic

process upon a fossil is often related to those which occurred early in the history. It has also been recognised that a taphonomic history may be determined by pre-mortem considerations, for example, the preference of certain habitats or diets (Shipman, 1981).

Most modern taphonomic studies focus upon the positive contributions that fossils and fossil assemblages can provide to our understanding of the biotic record, rather than being solely concerned with taphonomic losses. Seminal papers on this subject include works by Seilacher (1977), Behrensmeyer and Kidwell (1985), Thomas (1986), and Kidwell and Behrensmeyer (1988). Specific topics being researched include taphonomy's contribution to our knowledge of past ecosystems (e.g. Brett and Baird, 1986; Kidwell and Behrensmeyer, 1988) and depositional environments (Kidwell *et al.*, 1986; Behrensmeyer and Hook, 1992), the interaction of organisms and sediments (e.g. Kidwell and Aigner, 1983; Kidwell and Jablonski, 1983; Kidwell *et al.*, 1986; Brett and Baird, 1986; Brandt, 1989; Kidwell, 1991), time-averaging and relative rates of sedimentation (Behrensmeyer and Schindel, 1983; Kidwell and Bosence, 1991) and processes of fossilisation (papers in Allison and Briggs, 1991).

Many modern taphonomic studies rely heavily upon actualistic observational, experimental and analytical approaches. They also incorporate sedimentalogical and stratigraphic data, providing information upon depositional environment and burial profiles. Experimental studies have included analyses of mechanical rounding and sorting by transport in simulated fluvial or marine conditions (e.g. Voorhies, 1969; Wolff, 1973; Behrensmeyer, 1975, 1991; Hanson, 1980; Shipman, 1981; Allen, 1984, 1990; Argast *et al.*, 1987; Spicer, 1990). Other studies have included observations on decay of organisms under strict laboratory or closely monitored natural conditions (e.g. Dodson, 1973; Hill, 1979; Andrews and Cook, 1985). There has been a plethora of taphonomic theories based upon observations of hard-part modification by predation and scavenging (e.g. Dodson and Wexlar, 1979; Hill, 1979; Haynes, 1980a,b, 1982; Fisher, 1981; Andrews and Nesbit-Evans, 1983), biogenic corrosion (Brett and Baird, 1986; Brett, 1990), and the effects of weathering (Behrensmeyer, 1975; Korth, 1979) and trampling (Behrensmeyer *et al.*, 1986; Fiorillo, 1984, 1988b,c, 1989). Some of these studies (e.g. Fisher, 1981; Fiorillo, 1989) have incorporated experimental studies in order to demonstrate the application of their results to fossil concentrations. Understanding the processes of post-burial modification and eventually fossilisation has become a major theme of research in the 1990's, with experimental techniques being pioneered in understanding processes such as phosphatisation (e.g. Lucas and Prevot, 1991), pyritisation (e.g. Canfield and Raiswell, 1991b), charring (e.g. Cope, 1983), and calcitisation (Canfield and Raiswell, 1991a) of organisms.

The scope of some of these studies is often quite limited, taxonomically or in terms of the variables used to define a certain taphonomic parameter, and they can be rather simplistic in application of results from laboratory or controlled experiments to the

fossil record. For instance, much of the research concerning standardisation of disarticulation, transport and abrasion of vertebrate remains in fluvial regimes was carried out upon large mammalian skeletons and cannot be readily applied to accumulations of extinct non-mammalian microvertebrates, such as Hornsleasow. Others rely heavily upon the old theory that the present is truly the key to the past, for example understanding and recognition of modern predator activity is extremely useful, but how relevant is it to extinct faunas? The concepts and problems in applying actualistic procedures are further discussed in the relevant sections of Chapters 8 and 10, and are reviewed by Gifford (1981) and Lyman (1994).

That taphonomy as a science is thriving in the palaeontological community is indicated by the several books, conferences, special papers and review papers on the subject published in the last five to ten years. Some of the more important include Behrensmeyer and Kidwell (1988), Allison and Briggs (1991), Donovan (1990), Behrensmeyer *et al.* (1992), and Lyman (1994).

7.3. Basic principles in taphonomy

The two main taphonomic factors that a palaeobiologist needs to assess in looking at a fossil accumulation are:

(a) *The spatial and temporal derivation of the fossils*. In crude terms this is whether the fossils are *in situ* within the assemblage, or derived from elsewhere. Such an approach is applied to the assemblage as a whole (Kidwell *et al.*, 1986) and includes studying the taxonomic composition of the accumulation, any biofabric it might contain, the internal and external features of the concentration. Obviously the state of each individual fossil, as well as the structure and derivation of the assemblage as a whole, will reflect the sequence of events in their preservation, and components which exhibit different taphonomic histories can be sorted into grades which may reflect different derivations (Brandt, 1989). Johnson (1960) and Kidwell *et al.* (1986) defined six categories which can be applied to fossil assemblages: (i) *autochthonous or in situ assemblages*, in which the fossils are preserved in their original living positions; (ii) *parautochthonous or residual assemblages*, were fossils are still within their habitats but have been transported or reworked somewhat, and in which some organisms may have been removed; (iii) *allochthonous or transported assemblages*, in which fossils have been derived from another habitat; (iv) *remanie assemblages*, accumulations containing fossils reworked from significantly older deposits; (v) *piped assemblages*, were fossils are derived from younger deposits; and (vi) *mixed assemblages*, a combination of any of the above. These concepts are further described in a review of time-averaging (section 7.4.1) and summarised in Table 7.3 (after Kidwell and Bosence, 1991).

The genesis of the fossil assemblage should also be determined from a taphonomic study of these parameters. Kidwell *et al* (1986) identified three specific types

of genesis for fossil concentrations, which are: (i) *biogenic*; (ii) *sedimentologic*; and (iii) *diagenetic*. These are not mutually exclusive as, for instance, a biogenic accumulation may be concentrated even more by sedimentological forces or diagenetic controls upon the matrix in which it is preserved.

(b) *The taphonomic history of the specimen or assemblage.* The study of taphonomy is largely concerned with identifying the effects of post-mortem, pre-burial, burial and post-burial events upon faunal remains, which occur in a general chronological order known as the *'taphonomic history'* or *'taphonomic pathway'* (Lyman, 1994; Table 7.2.). From the death of the organism there are four specific stages of taphonomic history defined by Seilacher (1977) as: (i) *Necrolysis* - soft part decay and degradation; (ii) *Biostratinomy* - hard part modification, transport and sedimentation; (iii) *Fossilisation* - syndiagenesis of hard parts during burial; (iv) *Metamorphism* - including fossil metamorphism, late diagenesis and cements, and weathering. A fifth stage is the actual action of removing the fossil from its enclosing rock or sediment and conserving it for prosperity (Table 7.2). Table 7.2 illustrates a typical taphonomic history, in which the arrows signify the direction of the pathway from 'living community' to 'research collection', although in realistic taphonomic histories there may be several loops, for instance concerning reworking and re-burial.

The taphonomic history of a fossil can be resolved by analysing the traces or effects that have been left upon it by certain modifying processes and/or agents (Badgley, 1986). The traces of such events, such as tooth marks or abrasion, are known as *'taphonomic delimiters'* or *'taphonomic features'* (Table 7.2). The source of the modification, such as a river-system or predator, is known as the *'taphonomic agent'* and the action which changes the remains, such as being transported by water or bitten, is the *'taphonomic process'* (Lyman, 1994). Defining the taphonomic history of a fossil or accumulation provides the positive feedback to a case study, in releasing information upon the preservation of organisms in a loose chronological order. Therefore in this study I have attempted to describe features in terms of primary (occurring at death), secondary (preburial modifications) and tertiary (postburial) taphonomic processes (Metcalf, 1993).

The taphonomy has been studied separately for each of the three preserved organisms in the paleosols: the vertebrates; hard-shelled invertebrates; and plant fossils. The particular taphonomic procedures and goals for each group is slightly different, for instance, biological modifications such as tooth marks and acid digestion, are only preserved upon bone and shell material, whereas diagenetic modifications such as pyritisation are much more important in the non-vertebrate fauna. However, each group has essentially passed through a taphonomic path such as that shown in Table 7.1. Specifically in each group the procedure has been firstly to document the structure of the

ASSEMBLAGE	MODIFICATION	EXAMPLES	TAPHONOMIC DELIMITERS/FEATURES
LIVING COMMUNITY	AT DEATH	**1. DAMAGE TO REMAINS** (e.g. bite marks/gnaw marks; digestion) **2. ACCUMULATING EVENTS** (e.g. mass mortality)	1. punctures; grooves; scalloped edges; surface stripping. 2. assemblage characteristics as a whole
DEATH ASSEMBLAGE	SHORTLY AFTER DEATH (cadaver disarticulation)	**1. CADAVER TRANSPORT** **2. BIOLOGICAL EFFECTS** (e.g. trampling; decay; scavenger activity)	1. none 2. disarticulation; breakages (spiral fractures: 'SP'); surface scratching
BONE/SHELL ACCUMULATION	A. PREBURIAL	**1.SCATTERING/ CONCENTRATION EFFECTS** (e.g. transport) **2. SURFACE MODIFICATION** (e.g. weathering; breakages)	1. fossil assemblage as a whole 2. abrasion index (0-3); breakages (SP); weathering index (0-3); bone splitting: longtitudinal fractures: 'LO'
	B. DURING BURIAL	**1. SYNDIAGENESIS** (e.g. low pH groundwater dissolution) **2. PEDOGENIC/BIOTURBATION DISTURBANCE**	1. bones/shells pocked; early cements 2. breakages (SP)
FOSSIL ASSEMBLAGE	POST BURIAL/ DIAGENETIC	**1. DIAGENESIS** (e.g. mineralisation) **2. MOVEMENT** (e.g. compaction; deformation; exhumation)	1. late cements 2. breakage (transverse fractures: 'TR'); squashing/faulting; weathering
EXCAVATED SAMPLE	COLLECTION	**RECOVERY** (e.g. mechanical excavator damage; specimen loss)	breakage (TR)
RESEARCH COLLECTION	CONSERVATION	**1. SORTING/PREPARATION** (e.g. handling) **2. TREATMENTS** (e.g. acids)	1. brittle failure (crumbling); scratching 2. dulling/pitting of surface

Table 7.2. The taphonomic history and its effects upon a fossil or fossil acumulation: a living entity or community is changed into a research specimen or collection by a variety of modifying effects, which can be resolved from careful appraisal of the taphonomic delimiters of the particular specimen or sample. Vertical axis is time and arrows indicate direction of taphonomic pathway.

fossil assemblage as a whole, and then to describe the taphonomic history in approximate chronological order.

7.3.1. Taphonomic parameters for fossil assemblages

In determining the spatial and temporal relationships of a collection of fossils and the genesis of the accumulation, a taphonomist must study the structure and composition of the assemblage as a whole. The approach which I followed in ascertaining the taphonomic parameters of the Hornsleasow assemblage, was based upon the methods suggested by Shipman (1981), Brett and Baird (1986), Kidwell *et al.* (1986) and Badgley (1986), incorporating methodologies from both invertebrate and vertebrate taphonomy. This approach included considerations of the following parameters:

(a) *The structure of the accumulation.* This includes appraisal of the lateral and vertical distribution of fossils, their stratigraphic span, and disarticulation of specimens (Badgley, 1986). It also takes into account any inherent biofabric within the accumulation (e.g. root traces in position of growth), the general geometry and internal structure (e.g. size sorting) of the concentration (Kidwell *et al.*, 1986). A sedimentary description of the enclosing sediment can also be attempted at this point.

(b) *The biotic composition in the accumulation.* The first step in biotic analysis is to identify all the specimens in the assemblage. The specimens should be identified at two levels, as the element and taxon represented (Shipman, 1981). This composition of a particular accumulation depends upon the original palaeocommunity and taphonomic sorting of the assemblage, and hence, in describing this parameter information can be gleaned for both palaeoecological and taphonomic analyses. In terms of palaeoecological information which can be resolved, the faunal composition may elucidate taxonomic differentiation, the number of individuals per taxa, carnivore/herbivore ratios, and proportion of different age groups, and the sexes within an accumulation (Badgley, 1986; Pratt, 1979). Whilst taphonomic information can include the representation of the skeletal elements in the assemblage (Pratt, 1979), which can be compared to the proportions within the animal, or those within equivalent accumulations seen in natural and simulated environments of deposition (Behrensmeyer, 1975; Behrensmeyer *et al.*, 1979; Behrensmeyer & Dechant-Boaz, 1980; Bown & Kraus, 1981; Dodson, 1973; Voorhies, 1969) and those recovered from scatological accumulations (Dodson, 1973; Maas, 1985; Mellet, 1974; Korth, 1979). Even indeterminate fragments of bone (Shipman, 1981) or shell (Johnson, 1960) can be important in a taphonomic appraisal and therefore, in this analysis no fragment however incomplete or unidentifiable was discarded during sampling.

(c) *The condition of the specimens.* The general appearence of the specimens will be affected by physical or chemical damage occurring during the taphonomic history of the deposit (Table 7.2; Behrensmeyer, 1978; Dodson & Wexlar,

1979; Fiorillo, 1988b,c, 1991; Korth, 1979) and will be considered in detail within the following sections.

7.3.2. Primary taphonomic sorting

Primary taphonomic sorting is associated with the death of the individual organism or community by accidental, deliberate or natural causes (Metcalf, 1993). The first thing to realise about a collection of fossils is the obvious fact that fossilised remains are dead and this fact has a significant bearing upon the interpretation of a fossil assemblage (Hill, 1975). The mode of death of an organism or palaeocommunity can provide a considerable amount of taphonomic information. For example, loss or soft parts and the disarticulation of hard parts may occur at death. Hard-parts might be damaged, concentrated or even buried by the agent of death. Primary taphonomic sorting is complicated by the fact that some fossils, e.g. shed teeth or arthropod carapace moults, do not represent the remains of dead organisms, and hence dilute the taphonomic signature. Primary taphonomic sorting is particularly useful in providing good palaeoecological data, e.g. predator/prey relationships and preferences.

7.3.3. Secondary taphonomic sorting

Secondary taphonomic processes occur prior to the final burial and stabilisation of a fossil or assemblage (Metcalf, 1993). These include considerations of disarticulation, transport and fragmentation of skeletal remains. Fossil assemblage might be winnowed and their hard-parts modified, by the action of fluvial transport, flooding, weathering and the activity of scavengers, encrusters or bioeroders. These are important as they provide information concerning fossil/sediment interaction, the depositional environment in which the fossil is preserved or from which it has been derived, and prevalent palaeoclimatic conditions (Behrensmeyer, 1975; Brett and Baird, 1986). Finally, even the action of burial can produce a bias within the sample, as those fossils which are quickly stabilised by burial are, unless exhumed by erosion or bioturbation, removed from further modification by weathering or hydraulic transport (Metcalf, 1993).

7.3.4. Tertiary taphonomic sorting

These are the processes which normally occur after the organism has been buried (Metcalf, 1993). They include diagenetic alteration and cement precipitation, geochemical etching by soil or root acids, decomposition by the actions of biodegraders in the sediment, mechanical breakage caused by pedogenic, bioturbation or diagenetic stresses and compaction, and finally metamorphism. They are important in providing information on the conditions in which the fossil was preserved and upon the fossilisation process itself (Allison and Briggs, 1991). Lastly, it is conceivable that even within modern extraction and curation techniques, a sampling bias may be introduced (Shipman, 1981;

Metcalf, 1993). For example, less attractive and unidentifiable fossils might be discarded during curation. Breakages can, and frequently, do occur during collection and conservation of fossils, and hence it is also important to distinguish between these and earlier, taphonomically more important, fossil modifications.

7.4. Attempting palaeoecological reconstructions

A significant goal the palaeobiologist is to try to determine interrelationships within the fossil concentration, ideally the palaeoecology of the original sampled living community, or communities. The ultimate aim is to attempt to provide an idea of how the individuals of such a palaeocommunity interacted with each other and the surrounding palaeoenvironment (Behrensmeyer and Hook, 1992; Table 7.1). A further goal is to understand the individual lifestyles and habits of the organisms.

Fossilised remains cannot tell us everything about what an animal or plant was like in life, where it lived, what it interacted with and what was its preferred diet and habitat, although some of these features can be inferred from indirect evidence. A fossil populations consists of all organisms within the same interbreeding species, whereas palaeocommunities are made up of various populations which interact in a specific environment (Fagerstrom, 1964). Palaeocommunities such as fossilised reefs, dominated by firmly attached organisms are the most easiest to interpret (Fagerstrom, 1964), whilst communities made up of vagrant organisms, such as large vertebrates, commonly undergo a considerable amount of post-mortem alteration (Behrensmeyer *et al.*, 1979). In the latter assemblages, fossilised remains are not often associated with organisms with which they interacted in life, especially if they migrate or if the prevailing environmental conditions change over the duration of deposition. Establishing palaeoecological interaction is further complicated by the fact that organisms can be dispersed and modified after death, and the nature of their association in a assemblage might reflect the nature of the taphonomic agent rather than provide true palaeoecological information, for example, a carnivores' den accumulation will reflect the prey preferences, habit and habitat of the carnivore rather than those of the prey (Dodson, 1973; Mellet, 1974; Mayhew, 1977).

However, in most cases fossil assemblages are time-averaged accumulations dominated by indigenous material (Behrensmeyer *et al.*, 1979; Kidwell and Bosence, 1991). Not all animals and plants in a palaeocommunity are preserved, and the fossil assemblage recovered from Hornsleasow only provides a snapshot view of the life in and around the pool. For instance, although a few quartz-lined burrow traces of soft-bodied invertebrates such as earthworms have been found in the Hornsleasow Grey Clay paleosol, the structure of the clay matrix suggests that both soils were heavily bioturbated by a fauna that was not preserved. Nevertheless the concentration of such a fossil

accumulation can also provide evidence of the wider palaeoenvironment. Rare and transported elements can give an idea of the much wider floodplain palaeocommunity.

In general fossil assemblages can be sorted into several categories based upon the association of the fossils they contain, their individual or group taphonomic histories, and general features of the accumulation itself (Fagerstrom, 1964; Kidwell and Bosence, 1991). Criteria listed by Fagerstrom (1964) as important to elucidating the palaeoecology of a marine benthic community include an evaluation of the faunal composition and diversity, the size sorting and frequency of individual fossils, the ratio of disarticulated and articulated individuals, density of a population, ratio of fragmented to whole specimens, preservation status of fossils, along with more general details of the accumulation such as orientation of fossils and other sedimentary particles, and sedimentary textures, and evidence for post-burial alteration. Most of these criteria were already listed in the earlier sections on taphonomic evaluation of fossil assemblages and hence the two procedures are interlinked. An understanding of the taphonomic history of an assemblage can therefore provide a basis for interpreting how closely the accumulation approximates the original biotic associations (Fagerstrom, 1964). Palaeoecology is sorting and stripping away the taphonomic overprint which may have biased the fossil accumulation.

Kidwell and Bosence (1991, table 1) provide a summary of the terminology used in describing fossil concentrations and although this is mainly directed at marine benthic palaeocommunities, much of their work is applicable to this study. The terminology's used by these and other authors are listed in Table 7.3. Shipman (1982, p. 22) states that there are three general rules to adhere to when approaching a palaeoecological study of a fossil concentration and these are:

(1) Fossil accumulation ≠ biotic community.

(2) Fossils found in association did not necessarily interact in life.

(3) Abundance or absence of fossil(s) in the accumulation ≠ abundance or absence of individual(s) in living community.

The first rule is an obvious statement, but the second and third rule embrace the topics to be discussed in the next sections. These are the significance of time-averaging within a fossil accumulation, and the general assessment or estimation of relative or absolute abundances of individuals in the assemblage.

7.4.1. Time averaging

A fossil assemblage must take a certain period of time to accumulate, and it is during this period when time-averaging of the biota can occur. Time-averaged fossil accumulations were defined by Walker and Bambach (1971) as:

> "fossil assemblages .... that accumulate from the [local] live community during the time required to deposit the containing sediment".

Classification	Fagerstrom (1964)	Kidwell *et al.* (1986)	Kidwell & Bosence (1991)
I. Group of co-occurring live organisms			Community
II. Group of co-occurring dead organisms	Fossil assemblage	Fossil assemblage	Fossil assemblage
....that are thought to represent: A. Production by the local fauna ....and are thought to provide: 1. a nearly *perfect record of live* taxa with respect to species relative abundances, size-frequency distributions, and life positions or 2. a somewhat modified record owing to post-mortem disturbances and/or time-averaging of generations or to have been modified by 3. the addition of organisms from other habitats or by 4. the addition of organisms from later, ecologically distinct communities that occupy the site; community change may have been driven by physical environmental changes or by live/dead interactions or by 5. admixture with organisms of significantly greater age	 *Fossil census or* fossil community Residual census or residual fossil community Mixed fossil assemblage Mixed fossil assemblage Mixed fossil assemblage	 *Autochthonous* assemblage Parautochthonus assemblage Mixed parautochthon-ous) assemblage	Indigenous assemblage Census assemblage Within-habitat time-averaged assemblage Multi-habitat time-averaged or environmentally condensed assemblage Biostratigraphically condensed assemblage
B. Organisms that have been transported out of their life habitat	Transported fossil assemblage	Allochthonus assemblage	Allochthonous assemblage
C. Organisms reworked from significantly older deposits		Remanié assemblage	Remanié assemblage
D. Organisms introduced from significantly younger deposits.		Leaked assemblage	Piped assemblage

Table 7.3. Description and summary of classification of fossil accumulations after Kidwell & Bosence (1991).

Overall the fossil record is extremely incomplete, and this is partly as a result of the incompleteness of the sedimentary record. The fundamental control upon time-averaging is the rate and tempo of sedimentation, and the rough rule on time-averaging is that the longer that a deposit took to form, the more likely that taxonomic composition and the palaeoecology of the assemblage will be modified by differential preservation, mixing of successive generations and that the source, rate of supply and local environment of accumulation will change (Kidwell and Bosence, 1991). This is summarised in the classification of fossil assemblages given by Kidwell and Bosence (1991, table 1; Table 7.3).

Instantaneous mass-mortality 'census assemblages' are rare, as are those composed entirely of exotic organisms and in most cases, fossil assemblages are composed of time-averaged biotas, dominated by local material. These assemblages may form during periods in which a physical environment remained stable ('within-habitat assemblages') or during periods of environmental change in which the structure of the palaeocommunity changes as well ('palaeoenvironmental condensation': Fürsich, 1977). Table 7.3 and Kidwell & Bosence (1991) provide a good summary of the taphonomic parameters for each type of fossil assemblage.

Time-averaging is important because if deposition is slow, whilst biological rates are high, then assemblages will represent composite accumulations spanning decades or even hundreds of years, and any short-term palaeoecological information will be damped by long-term trends (Kidwell & Bosence, 1991). Even modern ecological studies have severe limitations when dealing with the interactions of a community over the span of biological time, and this is not nearly as long as geological time. There are three overall consequences of time-averaging to a palaeocommunity given by Kidwell & Bosence (1991), which are: (i) a heightened species richness; (ii) relative abundances that reflect long-term averages rather than successional states; and (iii) poor records of short-term ecological structure. Therefore in a fossil accumulation it is important not only to recognise that time-averaging has occurred, but to attempt to assess the degree of time-averaging which has taken place. This is considered by Kidwell & Bosence (1991, p.174) to be the "cornerstone of palaeoecological analysis".

To reconstruct the likely scale of time-averaging for a specific assemblage it is important to understand the dynamics of sediment accumulation, and the faunal composition and taphonomic history of the fossils themselves. Behrensmeyer and Schindel (1983) provide a useful guide to establishing the degree and magnitudes of time-averaging for several different palaeoenvironments, including terrestrial assemblages. Their paper documents approaches made by several authors using sedimentation rates, actualistic taphonomic analysis, geochemical and geophysical dating methods and morphological trend analysis. In this study only the former two approaches were used.

7.4.2. Counting individuals in a fossil assemblage

Generally, taphonomic histories tend to destroy information concerning the exact number of individual organisms interacting in a particular palaeocommunity, unless those organisms have not been transported or modified in any way (for example, the components of an ancient coral reef or a mass-mortality event). Although it may not be possible to establish the exact number of individuals within an assemblage, it is possible to assess the relative importance of a particular taxon within the original palaeoenvironment, and its role in the palaeocommunity. This last aim is not always possible to achieve, since ecological relationships can be difficult or impossible to establish, and taphonomic histories may overprint original relationships (Badgley, 1986; a good example would be the bias toward predators over prey in a natural trap, such as the Pleistocene Rancho La Brea tar pits in California; Sutcliffe, 1986).

Several methods of reviewing the numbers or relative abundances of whole individuals or populations in a fossil accumulation have been put forward, no single method being perfect for all assemblages (Lyman, 1994; Glinsky & Bennington, 1994). The most commonly used methods include:

(a) *Minimum number or individuals ('MNI').* This is defined by Badgley (1986) as the minimum number of individual organisms necessary to account for all the specimens counted. It is based upon counting the most abundant unique element (e.g. from one side of the body in vertebrates). This method works particularly well in a closely packed association of readily identified fossils, such as a benthic invertebrate community or mammalian vertebrate assemblage (e.g. Voorhies, 1969, Grayson, 1984), but has the significant problems of overrepresenting rare taxa and the subjective nature of determining associations between specimens (Badgley, 1986).

(b) *Number of identified specimens ('NISP') or maximum number of individuals.* This method identifies the relative abundance of a particular taxon by dividing the number of specimens (regardless of their uniqueness) by the total number of specimens in a sample. The method will always overestimate the abundance of a preserved taxon, regardless of its importance in the original palaeocommunity, but works well when studying sparsely fossiliferous palaeoenvironments such as river floodplains (Badgley, 1986).

(c) *Minimum number of skeletal elements per taxon ('MNE').* This method entails counting individuals based upon a nominated skeletal element for each taxon. All recognisable parts are counted however broken, and therefore this method works well in a closely associated assemblage which has become fragmented by taphonomic forces (Lyman, 1994). It has the advantage over the previous methods of giving a result intermediate between the maximum and minimum values of abundances.

(d) *Weighted abundance of elements ('WAE').* This method was proposed by Perkins (1964) and Holtzman (1979) and is based upon the frequency of elements per taxon divided by the number of preservable elements within the individual of that particular taxon. Holtzman (1979) argues that this method is more precise than MNI, especially in studying an invertebrate accumulation. However, Badgley (1986) points out that the method involves determining the number of skeletal elements which can be preserved in a sample, and for organisms with a large number of body parts (e.g. vertebrates) the method becomes unwieldy.

(e) *Relative frequency ('RF').* This was defined by Gilbert *et al.* (1982) as the total number of skeletal fragments per taxon corrected for the frequency of each skeletal element in the skeleton and the number of element types recovered for each taxon. This method incorporates parts of both MNI and NISP, and is considered to give a better representation of the biotic abundances by these authors. Again the method involves a prior knowledge of the skeletal proportions of an extinct organism, which may not be quantifiable.

For a good review of all these methods, their advantages and problems, the reader is directed toward Badgley (1986), Lyman (1994) and Glinsky & Bennington (1994).

In studying the Hornsleasow fossil assemblage every specimen no matter how badly fragmented was counted during taphonomic investigation. This kind of taphonomic investigation is converse to the laws which apply to determining palaeoecological interactions within supposed fossil communities; for example, methods of counting individuals of a given taxon within fossil assemblages rely on comparing and weighting different skeletal elements according to their abundance in the skeleton and distribution within the fossil sample (e.g. MNI or MNE). Much of this work, however, has been carried out upon large terrestrial mammalian communities (e.g. Shotwell, 1955; Badgley, 1986) or marine benthic invertebrate communities (e.g. Holtzman, 1979), and does not have any direct relevance for the palaeoecology of an extinct biota such as Hornsleasow. For example, the vertebrate community represented at Hornsleasow comprises extinct populations of reptiles, fishes, amphibians and few small mammals (Chapter 9). Palaeoecological studies have not been applied for the lower vertebrates or microvertebrate mammalian accumulations, and in the case of reptiles and fishes were an individual may release an indeterminate by high yield of scales, scutes and shed teeth, little can be done to assess the relative abundances of these populations. Therefore, no rigorous attempt has been made to establish the abundance of individuals or fossil populations for the Hornsleasow biota, although an estimation of the relative abundances has been attempted in most cases. These concepts are more closely considered in the relevant sections of the following chapters.

Chapter Eight Non-vertebrate biota

8.1. The Chipping Norton Formation biota

8.1.1. Hornsleasow

The paleosol clay horizons and overlying transgressive sand unit are the most fossiliferous part of the Chipping Norton Formation at Hornsleasow Quarry (Fig. 3.1, Tables 8.1-8.3; Appendix B2). Although the limestones have not been as carefully sampled as the clays, the impoverishment of macroinvertebrate remains, particularly in the oolitic units, is actual rather than a sampling bias. The upper biooosparites and biooomicrites of the 'Chipping Norton Member' (*sensu* Callomon *et al.*, 1993, Appendix B2) are slightly more fossiliferous than the oosparites of the 'Hook Norton Member' in the Hornsleasow sections. By far the most common identifiable macrofossil in all facies are oyster debris of the genus *Praeexogyra* (mostly *P. hebridica*) (Tables 8.1 and 8.2; Appendix B). In thin section, these limestones show very little identifiable bioclastic debris (up to 20% oyster debris in the more shelly facies) and microfossils. Poorly preserved gastropod casts and disarticulated echinoderm material are fairly common (Table 8.1). Cidarid spines and plates, asteroid ossicles and sometimes whole *Clypeus* have been recovered from the limestones surrounding the clay lens at Hornsleasow. Many of the gastropods found within the basal part of the clay lens have been reworked or washed into the hollow from the surrounding oolites (Table 8.2; Ashford, 1992). They include nerinids (*Nerinaea* sp.), genera of the family Pyramidellidae (*Eulima, Ceritherium*, and possibly *Chemnitzia*) and other small high-spired marine forms (Table 8.2; Ashford, 1992).

Trace fossils are rare except for clay-filled vertical and inclined *Skolithos*-type burrows which are fairly common in the upper beds of the succession. *Thalassinoides*, and *Rhizocorallium* traces have also been noted elsewhere in the formation in the northern Cotswolds (Sellwood & McKerrow, 1974). The formation of nested shell deposits and the heavily bioturbated beds at the top of the sequence suggests a lowering of energy and deposition by storm surges. This is borne out by the bored and encrusted hardground which caps the section at Hornsleasow and elsewhere in the northern Cotswolds (Sellwood & McKerrow, 1974). The paucity of macroinvertebrates in the Chipping Norton Formation at Hornsleasow has a bearing upon the nature of the deposition of the limestones, suggesting that the environment was somewhat restrictive to macrofauna.

UNIT	BIOTIC ASSEMBLAGE	ECOLOGY	TAPHONOMY	ABUNDANCE	REFERENCE
CNF	Foraminifera undifferentiated	M (B)	Fragments	Rare	S-74, H-92
CNM (HNM)	Bryozoa undifferentiated	M	Fragments	Fairly rare	S-74
CNF*	Serpulid undifferentiated	M	*In situ*, biodebris	Encrusters	S-74, M-p
	Brachiopoda				
CNF	*Kallirhynchia* sp.	M (epi.)	Casts; biodebris	Rare	S-74, H-87
CNF	*Rhactorhynchia obsoleta*	M (epi.)	Casts; biodebris	Rare	H-87
	Bivalvia				
CNF*	*Praeexogyra hebridica*	B-M (res. - epi.)	Biodebris; worn; whole; bored.	Hardground encrusters	S-74, H-87
CNM	*Lopha* sp.	M (epi.)	Biodebris	Rare	H-87
CNM	*Modiolus* sp.	M (epi.)	Biodebris	Rare	H-87
CNM	*Plicatula* sp.	M (epi.)	Biodebris	Rare	H-87
CNF	*Camptonectes rosimon*	M (mobile)	Whole - casts	Common	H-87
CNF	*Chlamys* sp.	M (mobile)	Whole -casts	Common	H-87
HNM (CNM)	*Astarte* sp.	M (in.)	Biodebris	Fairly rare	S-74,H-87
CNF	*Cucullaea* sp.	M (in.)	Biodebris	Rare	H-87
CNF	*Entolium* sp.	M (in.)	Biodebris	Rare	H-87
HNM	'*Lucina*' sp.	M (in.)	Biodebris	Common	S-74
CNM (HNM)*	*Myophorella signata*	M (in.)	*In situ*/biodebris	Fairly rare	H-87, M-p
CNM (HNM)*	*Plagiostoma cardiiformis*	M (in.)	*In situ*/biodebris	Fairly rare	H-87, M-p
CNM (HNM)	*Pleuromya* sp.	M (in.)	Biodebris	Fairly rare	H-87
CNM*	*Trigonia* sp.	M (in.)	Casts; biodebris	Rare	H-87, V-89
CNM	*Vaugonia moretoni*	M (in.)	Casts; biodebris	Rare	H-87
	Gastropoda				
CNF*	High-spired 'nerinid'-type	M (epi.)	Casts; few whole	Common	S-74, M-92
CNM*	*Bathonella (Viviparus)* sp.	FW-B	Whole-casts	Rare (FW lenses)	S-74, M-92
CNM*	*Nerinaea* sp.	M	Casts; few whole	Rare	A-92
CNM*	*?Ceritella* sp.	M	Casts	Common	A-92
CNM*	*?Eulima* sp.	M	Casts	Rare	A-92
CNM*	*Cerithenum* sp.	M-B	Casts; few whole	Fairly rare	A-92
	Cephalopoda				
CNM	*Oppelia (Oxycerites) limosa*	M		Very rare	C-80, H-87
HNM	*Zigzagiceras (Procerozigzag) pseudoprocerus*	M		Very rare	C-80
	Echinodermata				
CNF	*Clypeus mulleri*	M	Casts; few whole	Fairly common	H-87, M-p
CNF	Cidarid spines	M	Fragments; worn	Common	H-87, M-p
CNF	Isocrinid ossicles	M	Fragments; worn	Common	M-p, A-92
	Chondrichthyes (Appendix A2)				
CNM	*Asteracanthus* sp.	M	Teeth	Rare	M-p
CNF	*Hybodus* sp.	M	Teeth	Fairly rare	S-74, H-87
	Osteichthyes (Appendix A2)				
CNM (HNM)	Pycnodont-form teeth	M	Teeth	Fairly rare	M-p
CNF	Fish scales	?M	Scales	Rare	M-p
	Reptilia (Appendix A2)				
?HNM	Testudines undifferentiated	?M	Postcranial	Very rare	G-83
CNF	*Teleosaurus* sp.	M	Undifferentiated	Very rare	R-39
CNF	*Steneosaurus* sp.	M	Undifferentiated	Very rare	R-39
HNM (CNM*)	*Cetiosaurus oxonensis*	T	Undifferentiated	Very rare	B-78, R-11, R-29
CNF	*Megalosaurus bucklandi*	T (coastal)	Teeth	Very rare	R-29, G-83
?HNM (CNM*)	*Lexovisaurus vetustus*	T	Postcranial	Very rare	G-83
	Trace fossils				
CNM*	*Diplocraterion* sp.	M		Very rare	M-p
CNM*	*Rhizocorallium* sp.	M		Rare	S-74
CNF*	*Skolithos* sp.	M (res.)	Clay-filled	Common	S-74, H-87, M-p
CNM (HNM*)	*Thalassinoides* sp.	M		Fairly rare	M-p
	Plants				
HNM (CNM*)	Wood material (?conifer)	T (upland)	Lignite/coalified	Fairly common	S-74, M-p
HNM (CNM*)	Charcoal	T	Fragments/specks	Fairly common	M-p

Table 8.1. Biotic assemblage for different facies of the Chipping Norton Limestone Formation (CNF), including the limestones of the the Chipping Norton Member (CNM) and Hook Norton Member (HNM). The ecology (ECLGY) of each taxa is also listed, and the abbreviations are as follows: (1) for the invertebrates 'M' marine, 'M(in.)' infaunal, 'M(epi.)' epifaunal, 'B' brackish, 'FW' freshwater; (2) for the floral assemblage: 'A(fw-olg)' aquatic (freshwater-oligotrophic), 'T-bog' terrestrial boggy or marshland flora, 'T-upland' upland flora, 'T-SA' semiaquatic bog-pond flora; (3) for the vertebrates: 'A(c)' cosmopolitan aquatic, 'SA(fw)' freshwater semiaquatic, 'A(m)' marine aquatic, 'T' terrestrial. Reference abbreviations are as follows: 'A-92' Ashford, 1992; 'B-77' Beeseley, 1877; 'C-50' Channon, 1950; 'C-80' Cope *et al.*, 1980; 'C-89' Cole, 1989; 'G-74' Galton, 1983; 'H-91' Haslett, 1991; 'H-92' Haslett, pers. comm. 1992; 'H-87' Horton *et al.*, 1987; 'M-p' pers. obs.; 'M-92' Metcalf *et al.*, 1992; 'P-78' Phillips, 1878; 'R-11' Richardson, 1911; Richardson, 1929; Richardson *et al.*, 1933; 'S-74' Sellwood & McKerrow, 1974; 'T-69' Torrens, 1969; 'V-89' Vaughan, 1989; 'W-94' Metcalf & Walker, 1994.

8.1.2. Other localities

Elsewhere the Chipping Norton Formation has seldom yielded well-preserved macrofossils (Table 8.1). The fauna is dominated by shallow marine bivalve taxa, particularly thick-shelled sessile epifaunal forms such as the oyster *Praeexogyra* (formerly *Liostrea*), which makes up most of the biodebris within the oolites (Horton *et al.*, 1987) and occasionally occurs in bed-forming concentrations and *in situ* encrusting hardgrounds (e.g. at the top of the 'Chipping Norton Member' at Hornsleasow; Fig. 3.1). Other epifaunal forms are *Lopha*, *Modiolus* and *Plicatula*. Mobile species are also fairly abundant, including *Camptonectes* and *Chlamys*. Less common are the shallow infaunal bivalve species which include *Astarte*, *Cucullaea, Entolium*, *Myophorella*, *Plagiostoma*, *Pleuromya* and *Trigonia* (Horton *et al.*, 1987) and are rarely found in life position. Other invertebrates are much rarer although echinoderms, gastropods, brachiopods, bryozoa, foraminifera and serpulids have been found (Sellwood & McKerrow, 1974). Rare trace fossils include *Skolithos*, *Thalassinoides, Rhizocorallium* and *Diplocraterion* (Sellwood & McKerrow, 1974). Ammonites are extremely uncommon, complicating correlation (section 2.5.2). Plant material, fish teeth and bone material make up a low, but significant proportion of the assemblages (section 2.2). Plant remains suggest a close proximity to land and occur as large pieces of lignified wood or disseminated 'black specks' of charcoalled 'ash', which are frequently observed within sections and hand specimen.

The Swerford Member marginal-facies in north-east and eastern Oxfordshire contain large amounts of lignite, associated with fresh-water invertebrates and suggest low salinities and rapid rates of sedimentation from the nearby coastal marshes (section 3.5; Sellwood & McKerrow, 1974). The fauna and lithologies of the Chipping Norton Formation *sensu stricto* in central and north-west Oxfordshire are suggestive of extremely shallow marine conditions, accumulating close to a shoreline north of the area. The original environment seems to have been one of moderate energy, with a subsequent increase in current activity over time. Most of the biodebris consists of rare oysters and thick-shelled pectinids which are concentrated in lag deposits. The valves are disarticulated, fairly abraded, and sometimes nested, suggesting formation by contemporaneous winnowing and re-distribution of sediment during storms (Sellwood & McKerrow, 1974). The storm-lags are probably formed by a combination of unstable substrate conditions with continuous abrasion and reworking, and probably also account for the mixing of ooliths and coated and non-coated sand and skeletal grains.

In north-west Oxfordshire, the Chipping Norton Formation environmental conditions were similar to the carbonate sand-belts of Florida and the Bahamas (Chapter 6). The paucity of macrofauna and its lack of diversity, plus the rarity of burrowing organisms, also point to unstable substrate conditions and high environmental stresses

(Horton *et al.*, 1987). Fresh-water deposits at Sarsden and Sharps Hill (section 3.5), containing wood, charophytes and the gastropod *Bathonella* (*Viviparus*), also attest to the extremely shallow and unstable nature of depositional conditions of the Chipping Norton Formation. Nowhere is this more true than for the limestones of Hornsleasow and the other sections on the flanks of the Moreton-in-Marsh swell structure (section 2.6.2; Sellwood & McKerrow, 1974). Here the faunal assemblages are similar to the fragile ecologies of the unstable substrates on the Bahamas Bank (cf. Coogan, 1971).

8.2. The Hornsleasow paleosol biota

To date, the Hornsleasow paleosol units have produced a spectacular diversity of vertebrate fauna from a variety of different terrestrial and aquatic ecological niches (Chapter 9). Other *in situ* finds include plant material, freshwater gastropods, ostracods, and coprolites (Tables 8.2 and 8.3). Much derived fragmentary marine invertebrate material was also recovered from the clays. The karst layer and the laterally-impersistent clay lens are overlain by the marine carbonate sand unit ('Transgressive Layer') containing bivalve and echinoid debris, sharks' teeth, dinocysts, and foraminiferans.

8.2.1. The invertebrate fauna and palaeoecology (Table 8.2)

The invertebrate assemblage within the clays comprises two main components, a derived accumulation of marine forms and an *in situ* fauna of fresh-water forms (Fig. 8.1a).

The first component consists of poorly preserved biodebris, making up around 40% of the assemblage. The majority (60-70%) of this being fragmental oyster shell debris (*Praeexogyra* and *Lopha*), and other indeterminate bivalve fragments (Table 8.2). Recognisable fragments of the large pectinid *Plagiostoma* have also been recovered (Ashford, 1992). Echinoid debris is also fairly common (Fig. 8.2a,b), and includes extremely abraded isocrinid ossicles which often have ooliths adhering to the surface (Fig. 8.2c,d). High-spired 'nerinid'-type gastropod casts are fairly abundant within the normal and marginal marine facies of the Great Oolite Group and many of these forms also occur within the clays (Fig. 8.2e). A student at Bristol University (Ms. S. Ashford) identified some of these under my supervision for her BSc project (Ashford, 1992) and her results are given in Table 8.2. Her identification of the Great Oolite Group genera *Nerinaea* and the ceritherid *Ceritherium* appear to be correct (Table 8.2), but the genera *Ceritella*, *Eulima* and *Chemnitzia* were only defined upon fairly degraded and broken single specimens, and on re-examination of the material, I consider these identifications to be suspect. The low-spired globuse marine 'naticid'-forms are also present, and the genera *Natica* and *Trochus* (Fig. 8.2f) were identified by Ashford (1992). These classifications should be again treated with a degree of caution, as both genera were named upon poorly preserved specimens (Fig. 8.2f) and gastropods of the freshwater family Viviparidae can also be 'naticiform'. Nevertheless, there appears to be a large

UNIT	BIOTIC ASSEMBLAGE	ECLGY	TAPHONOMIC ASSEMBLAGE	DERIVATION (in clays)	REFERENCE
	Foraminifera				
Gy/Gn	*Epistomina regularis*	M-B	Well preserved	storm deposit	H-91, H-92, M-92
Gy/Gn	*Trochammina canningensis*	M	Well preserved	storm deposit	H-91, H-92, M-92
Gn	*Ammobaculites cf. agglutinans*	M	Well preserved	storm deposit	H-91, H-92, M-92
Gn	*Harplophragmoides canui*	M	Well preserved	storm deposit	H-91, H-92, M-92
Gn	*Paalzowella fiefeli*	M	Well preserved	storm deposit	H-91, H-92, M-92
Gy/Gn	Foraminifera Gen. et sp. indet.	M	Poorly preserved	reworked - CNF	H-91, H-92, M-92
Gy/Gn	Ostracoda Gen. et sp. indet.	FW	Well preserved	*in situ*	A-92, H-92, M-92
	Bivalvia				
Gy/Gn	*Praeexogyra hebridica*	B-M	Fragments; worn	storm; reworked - CNF	C-50, T-69, M-92
Gn/Tr	*Plagiostoma cardiiformis*	M	Whole	storm; reworked - CNF	C-50, T-69, M-92
Tr	*Trigonia* sp.	M	Whole; fragments	n/a	T-69, M-93
	Gastropoda				
Gy/Gn	*Viviparus cf. scoticus*	FW (B)	Well preserved	*in situ*	M-92
Gy/Gn	*Vivaparus cf. bernicia*	FW(B)	Well preserved	*in situ*	A-92
Gy/Gn	*Valvata cf. comes*	FW-B	Well preserved	*in situ*	A-92, M-92
Gy/Gn/Tr	Naticidae Gen. et sp. indet.	M-B	Casts; few whole	storm deposit	A-92
Gy/Gn	*Natica ?globosa*	M-B	Casts	storm deposit	A-92
Gy	*?Trochus* sp.	M	Cast	reworked - CG	A-92
(Gy) Gn	Numerous high-spired 'Nerinid'-type gastropoda	M	Casts; few well preserved	reworked - IO/CNF; ?storm deposit	A-92, M-92 A-92
Gy/Gn	*Nerinaea* sp.	M	Casts; few whole	reworked - ?CNF	A-92
Gy (Gn)	*?Ceritella* sp.	M-B	Casts - common	reworked - CNF	A-92
Gy	*?Eulima* sp.	M	Cast	reworked - CNF	A-92
Gy	*Cerithenium* sp.	M-B	Casts; few whole	reworked - CNF	A-92
	Echinodermata				
Gy/ Gn	Cidaroid spines	M	Fragments; worn	storm deposit/reworked	H-91, A-92, M-92
Gy	Isocrinid ossicles	M	Fragments; worn	reworked - CNF	A-92

Table 8.2. Invertebrate assemblage recovered from the Hornsleasow clays, showing palaeoecology, taphonomy and possible derivation. Abbreviations as for Table 8.1. except for 'Gy' Grey Clay, 'Gn' Green Clay, 'Tr' transgressive unit, 'IO' Inferior Oolite and 'CG' Clypeus Grit.

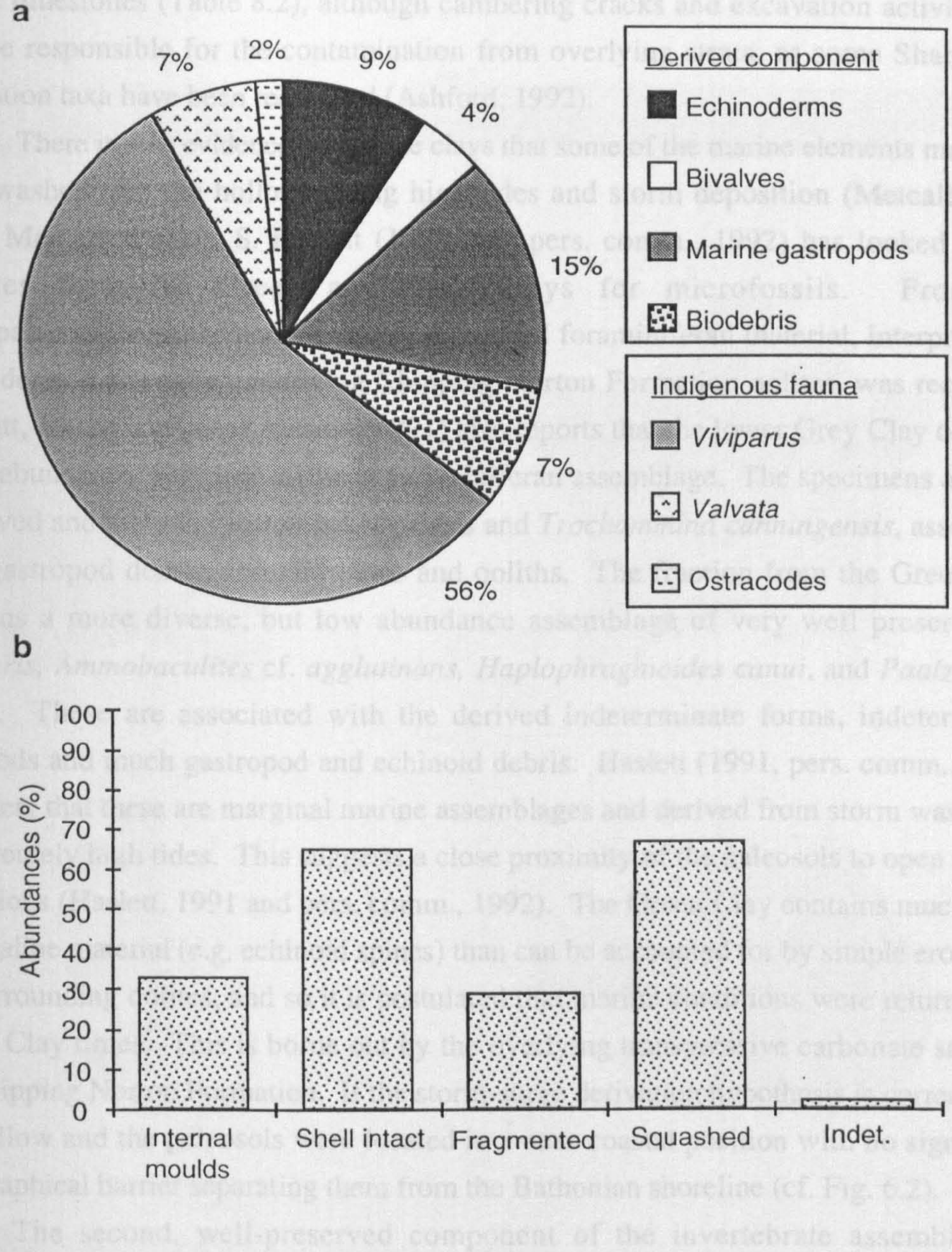


Figure 8.1 Invertebrate faunal composition (a) by percentage in sieved fractions. Taphonomy (b) showing the percentage of the indigenous gastropod shells preserved as internal moulds, with or without intact shell, and the percentage of those fragmented or compressed during fossilisation.

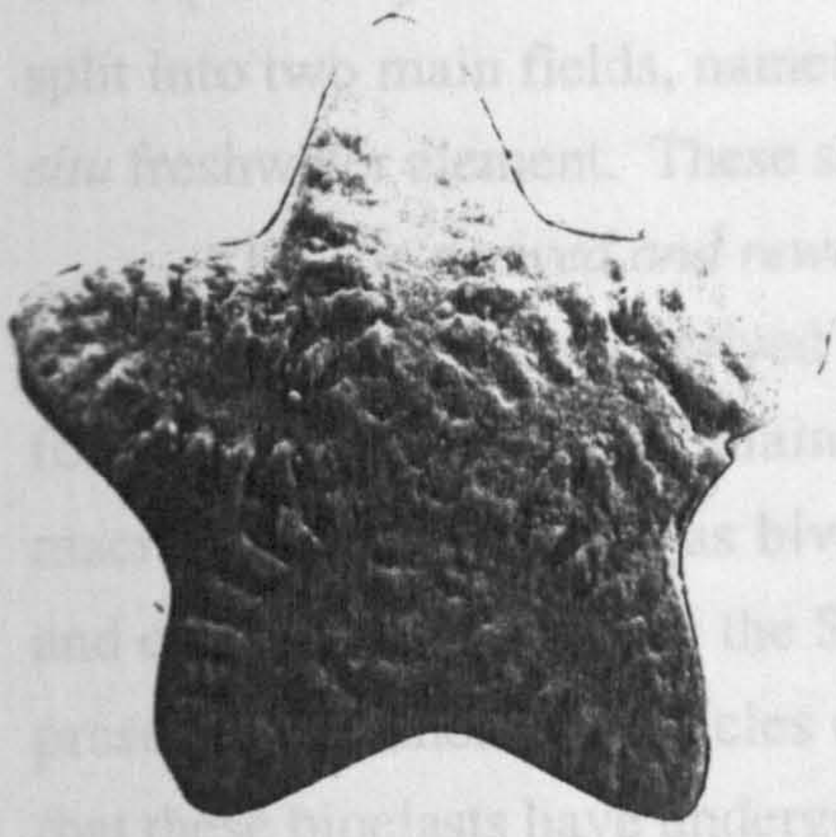
proportion of these marine forms within the clays (Table 8.2). Clearly this element of the fauna is derived, mostly from the underlying Chipping Norton Formation and Inferior Oolite limestones (Table 8.2), although cambering cracks and excavation activity may also be responsible for the contamination from overlying strata, as some Sharps Hill Formation taxa have been recovered (Ashford, 1992).

There is also evidence within the clays that some of the marine elements may have been washed into the hollow during high-tides and storm deposition (Metcalf *et al.*, 1992, Metcalf, 1994). S. Haslett (1991, and pers. comm., 1992) has looked at two samples from the Green and Grey Clays for microfossils. From the micropalaeontological fractions poorly preserved foraminiferan material, interpreted as being derived from the underlying Chipping Norton Formation oolites, was recovered (Haslett, 1991). However, Haslett (1991) also reports that the lower Grey Clay contains a low-abundance, very low diversity foraminiferan assemblage. The specimens are well preserved and include *Epistomina regularis* and *Trochammina canningensis*, associated with gastropod debris, serpulid tubes and ooliths. The fraction from the Green Clay contains a more diverse, but low abundance assemblage of very well preserved *E. regularis, Ammobaculites* cf. *agglutinans, Haplophragmoides canui*, and *Paalzowella fiefeli*. These are associated with the derived indeterminate forms, indeterminate ostracods and much gastropod and echinoid debris. Haslett (1991, pers. comm., 1992) considers that these are marginal marine assemblages and derived from storm wash-over or extremely high tides. This suggests a close proximity of the paleosols to open marine conditions (Haslett, 1991 and pers. comm., 1992). The Green Clay contains much more stenohaline material (e.g. echinoid spines) than can be accounted for by simple erosion of the surrounding oolites, and so it is postulated that marine conditions were returning by Green Clay times. This is borne out by the overlying transgressive carbonate sands of the Chipping Norton Formation. If the storm-surge derivation hypothesis is correct, then the hollow and the paleosols were located in a near-coastal position with no significant topographical barrier separating them from the Bathonian shoreline (cf. Fig. 6.2).

The second, well-preserved component of the invertebrate assemblage is essentially non-marine (Fig. 8.1a), and consists of planispiral gastropods (Figs. 8.3b-g and 8.4), and some rare ostracods (Fig. 8.3a). These are considered to be autochthonous. R.G. Clements (pers. comm., 1991) has identified the gastropods as *Viviparus* cf. *scoticus* (Tate) and *Valvata* cf. *comes* (Hudleston). The former is tentatively regarded as non-marine, and tolerant of low-salinity, possibly fresh-water to oligohaline (brackish) conditions (McKerrow & Kennedy, 1973). By comparison with Upper Jurassic species of the genus, *Valvata* cf. *comes* could tolerate oligohaline to polyhaline conditions. On balance, the salinity in which the gastropods were living can be regarded as low to moderate, and most likely the former (R.G. Clements, pers. comm., 1991). These gastropods, and in particular *Viviparus*, are extremely abundant

a

b



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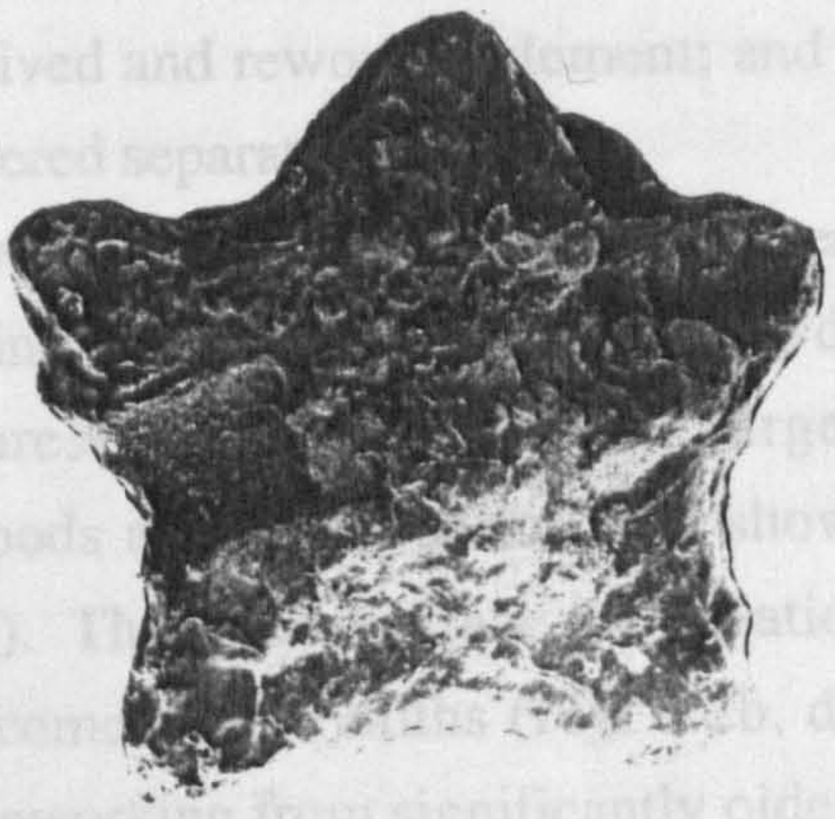

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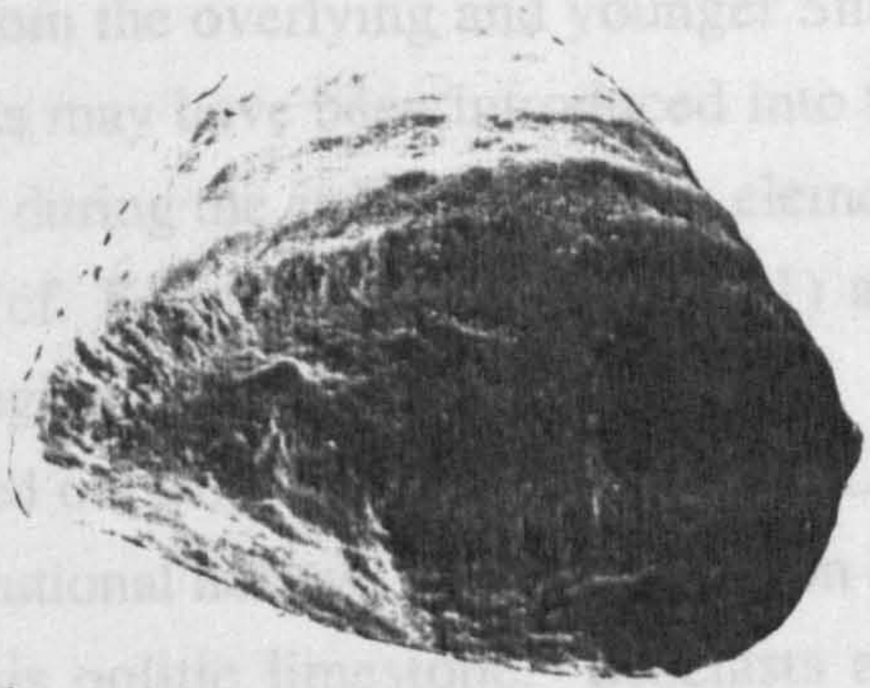


f

Figure 8.2. Reworked marine invertebrates from the underlying Chipping Norton Formation limestones : (a) G. 51322, cidaroid spine (10.8mm in length); (b) G. 52760, cidaroid spine (7.5mm in length); (c) G. 52757 and (d) G. 54807, isocrinid ossicles (both 4.2mm in maximum width); (e) G. 89777, 'nerinid'-type gastropod (9.8mm in spire height); (f) G. 52763, *Trochus* sp. (3.8mm in spire height).

within the lower Grey Clays supporting the sedimentological evidence that this horizon represented deposition within a fresh-water pool (Chapters 4-5). They are consequently much rarer within the dryer sediments of the upper Green Clay horizon.

8.2.2. Invertebrate taphonomy

The Hornsleasow invertebrate fauna can essentially be described as a 'mixed' assemblage (*sensu* Kidwell & Bosence, 1991), with an indigenous and an allochthonous component. The taphonomy of the invertebrates recovered from the Hornsleasow paleosols can be split into two main fields, namely: (1) the derived and reworked element; and (2) the *in situ* freshwater element. These shall be considered separately below.

(1) *The derived and reworked invertebrate fauna.* The reworked elements within the paleosols are largely derived from weathering of the underlying limestones during soil formation. They are mainly poorly preserved fragments of larger robust macroinvertebrates such as bivalves, gastropods and echinoderms, and show abraded and eroded surfaces under the SEM (Fig. 8.2). Their poor surface preservation and the presence of adhering particles of limestone cement and ooliths (Fig. 8.2b, d), suggest that these bioclasts have undergone physical reworking from significantly older deposits. They are thus termed 'remanie' assemblages (cf. Kidwell & Bosence, 1991). There is some evidence that a few specimens came from the overlying and younger Sharps Hill Formation (Ashford, 1992). These specimens may have been introduced into the clays by cambering cracks within the quarry face or during the excavation. This element of the fauna is known as the 'piped' assemblage (cf. Kidwell & Bosence, 1991) and is an important consideration in terms of time-averaging.

As these elements have been weathered out from the Chipping Norton Limestone which underlies and surrounds the karstic solutional hollow, their composition is similar to bioclasts seen within thin sections of this oolitic limestone. Bioclasts are either composed of a replacement low magnesium calcite which preserves original textures (both in hand specimen and thin section), or they have suffered carbonate dissolution and secondary growth of the yellow-orange ferroan calcite. All this mineralisation occurred prior to reworking. The badly corroded and pitted surface preservation of some of the bioclasts as revealed by the SEM (e.g. Fig. 8.2a,c,e,f) is attributed to dissolution of the calcite by chemical weathering during karstification of the sub-soil limestones. Humic and other organic acids are largely responsible for carbonate dissolution within soils, and bioclasts in close proximity to decaying organic material would be prone to corrosion.

The derived or exotic forms interpreted by Haslett (1991) as being deposited within the karstic hollow during storm wash-over and high tides are essentially well-preserved marine biodebris and microfossils. They are usually disarticulated, and often broken, but are not significantly abraded or corroded (Haslett, 1991). They form an allochthonous assemblage of out of habitat faunal communities (*sensu* Kidwell &

Bosence, 1991). Therefore, their presence within the hollow can offer no further palaeoecological insight into the conditions within the pond, except that the position of the paleosols was coastal.

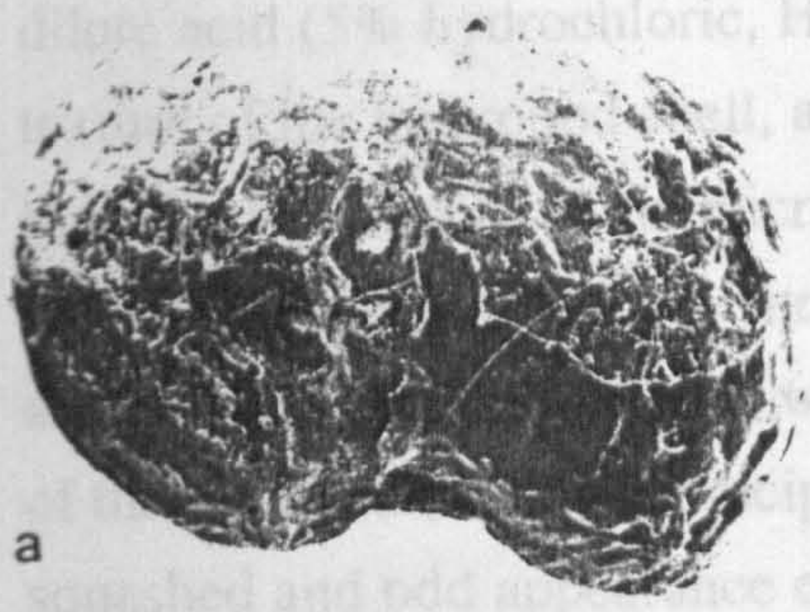
(2) *The indigenous freshwater invertebrate fauna.* This assemblage is characterised by the two gastropods, *Valvata* and *Viviparus*, of which the latter is by far the most abundant (Fig. 8.1a). It is essentially an indigenous fauna, although one that has been exposed to post-mortem and post-burial modifications (for example, fragmentation and compaction: Fig. 8.1b) and time-averaging, and hence, does not represent an *in situ* fossil community (*sensu* Jarvinen *et al*, 1986). Kidwell & Bosence (1991) have suggested the term "within-habitat time-averaged assemblage" for such a concentration of fossils.

In general, the features of the indigenous invertebrate fauna within the Hornsleasow clays are those of a low-diversity and low-abundance epifaunal community (Kidwell, 1991, her table 2, p.218). However, in some concentrations within the Grey Clay individuals of both genera are found in relative abundance, and may constitute up to 60% of the organic clasts recovered from sieving (R. Walker, unpublished data, 1992 - organic clasts make up less than one per cent of the volume of sediment washed). The assemblage has a mixed size, and presumably, age spectrum with specimens ranging between one and twenty millimetres (Figs. 8.3. and 8.4). The shells are also fairly well dispersed throughout the clays and show no preferred orientation within the sediments. Around 30-40% of the bioclasts are broken or show evidence of damage. All these factors suggest that the assemblage has become time-averaged, probably by bioturbation and pedogenic effects.

None of the Hornsleasow gastropods exhibit any trace of their original aragonite mineralogy. The shells within this assemblage display three modes of preservation:

(i) *Calcitised bioclasts.* The first mode of preservation occurs only in the small specimens (between 1-5mm). In these, the aragonitic shell has been completely replaced by the more stable calcite crystal configuration, in an early diagenetic, near-surface process known as "calcitisation" (Bathurst, 1975). This process is not the same as the wholesale dissolution and late diagenetic void-filling drusy calcite replacement of shells which is often seen in carbonate rocks (e.g. those preserved as ferroan-rich calcite within the Chipping Norton Formation limestones), but has occurred by dissolution and reprecipitation of the shell calcium carbonate across thin reaction front (Tucker, 1990). The crystals of calcite are quite coarse (Fig. 8.3e), but preserve fine anatomical details of the original shell, such as growth banding (Figs. 8.2b,f and 8.4e). These calcitised shells are also infilled with clay or more rarely with framboidal or aggregated pyrite (Fig. 8.3g,h).

(ii) *Haematitic and limonitic casts.* The second mode of preservation is as haematitic and limonitic casts. These show a cracked and extremely fine 'dusting' of



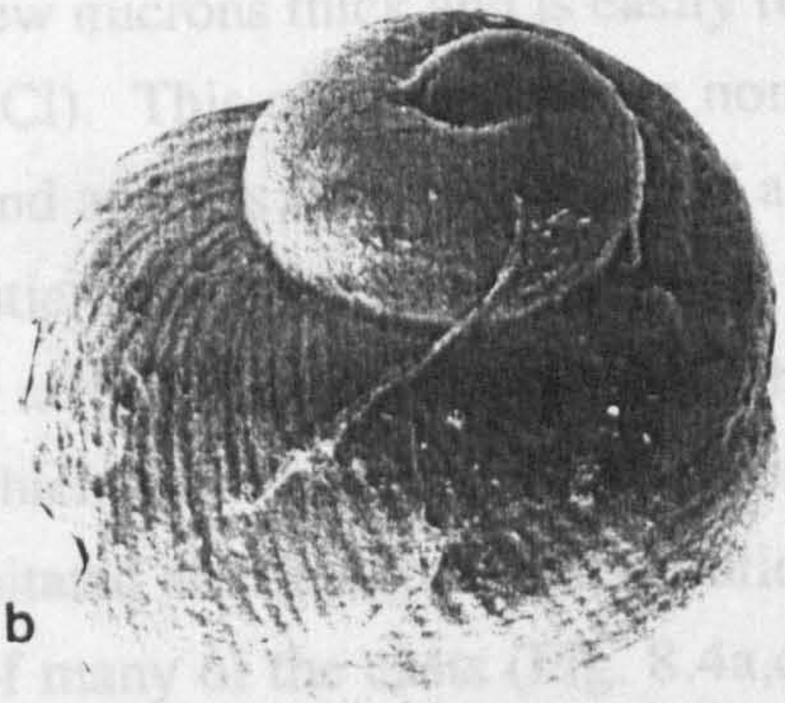


Figure 8.3. Indigenous invertebrate material from the Hornsleasow clays : (a) G. 50127, ostracod (1.4mm diameter); (b-g) viviparid freshwater gastropods, (b) G. 54038, calcitised gastropod (1.3mm width); (c) G. 56901, calcitised gastropod (1.1mm width); (d-e) G. 89600, calcitised gastropod (4.4mm width) and inset (e) calcite crystals (10mm = 4μm); (f) G. 56342, broken gastropod (1.7mm aperture diameter); (g-h) G. 60901, pyritised gastropod (1.3mm width) and inset (h) aggregated pyrite (10mm = 10μm).

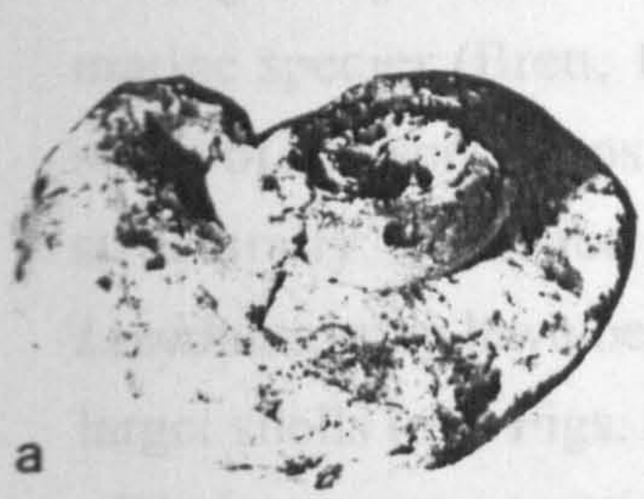
surface calcite, which is only a few microns thick and is easily removed by brushing with dilute acid (5% hydrochloric, HCl). This coating displays none of the original surface texture of the gastropod shell, and appears to be a residue of a micritic envelope rather than the shell itself. This micritic envelope represents a thin zone of alteration of the aragonite shell produced by the action of endolithic algae. The red-brown iron oxides and yellow-brown hydroxides which infill the mould exhibit none of the original textures of the shell and may have precipitated after aragonitic dissolution: this would explain the squashed and odd appearance of many of the casts (Fig. 8.4a,c; Clements pers. comm., 1991). It is likely that the early diagenetic minerals were precipitated from the dissolved ferrous iron in much the same way as the ferruginous rhizocretions which surround many of the rootlets in the Green Clay (section 5.4.3).

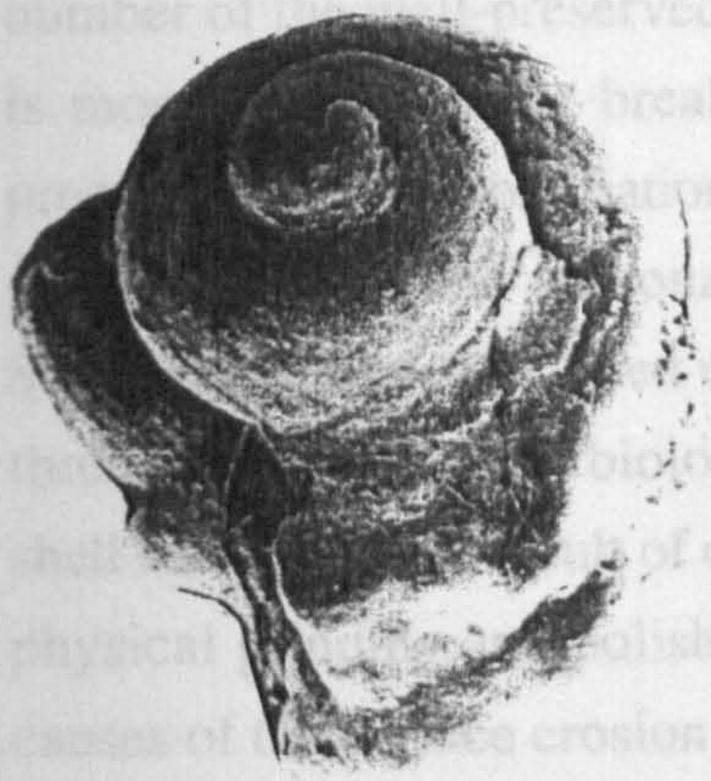
(3) *Phosphatised internal moulds.* The third mode of preservation is as internal moulds within phosphatic pedogenetic nodules within the Grey Clay. Within these, the shell has been entirely dissolved away during formation of the nodule, leaving a perfect mould exhibiting fine detail of the original shell-shape and structural details, such as growth banding and ribbing. It appears that the phosphate nodules have nucleated around the gastropod shell (section 5.9.2). In other studies of phosphatised shell material, the apatite is seen to replace the calcium carbonate to produce perfectly preserved casts of the shell (Prevot & Lucas, 1990). However, in the Hornsleasow nodules, there may have been an excess of phosphate available within the sediment and hence, only moulds of the shells are seen in freshly split nodules.

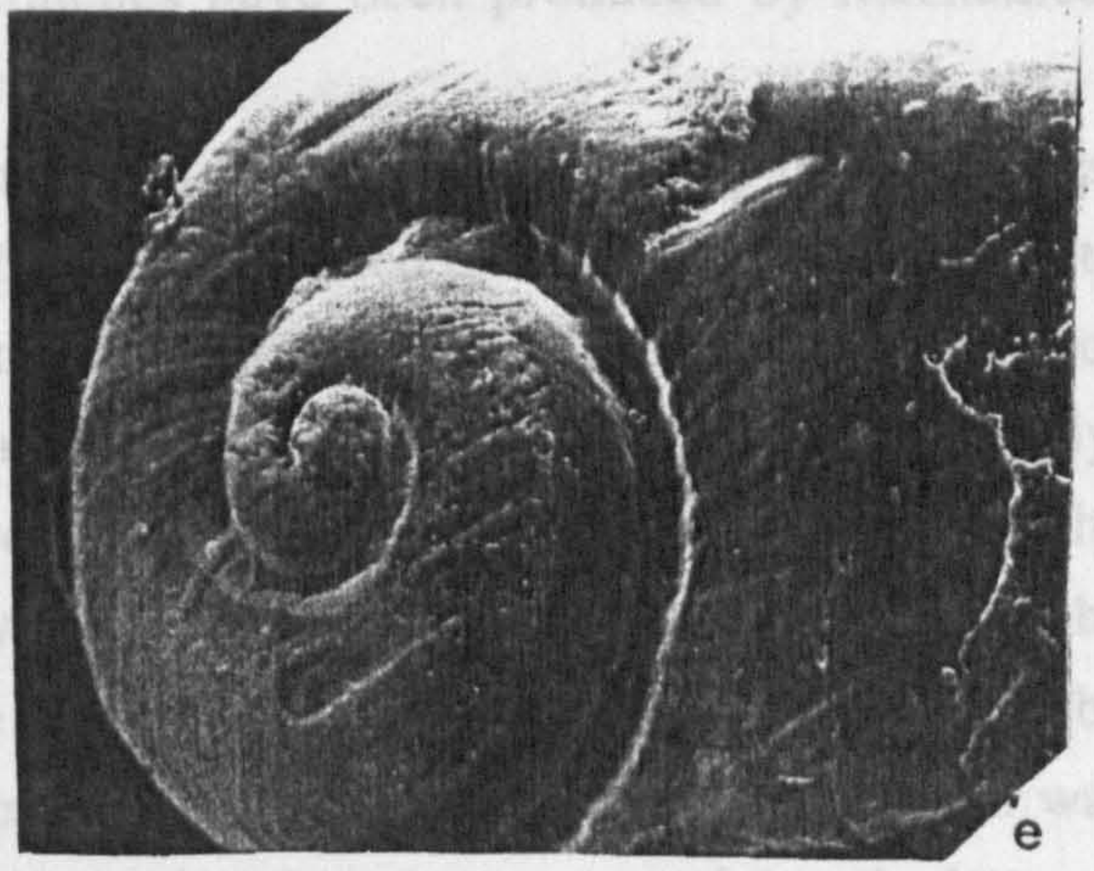
8.2.3. Discussion

Shells with identical preservational mode are often found together within the sieved fractions, suggesting that microenvironmental taphonomic conditions were responsible. Local clusters of shells can produce favourable pore-water conditions for carbonate preservation even within an acidic soil (Kidwell & Bosence, 1991). This is facilitated by the partial dissolution of biogenic carbonate, and the build-up of the bicarbonate (HCO^{3-}) anion, which buffers the acidic conditions (Canfield & Raiswell, 1991a). Aragonitic shells may be altered to calcite by early post-mortem modification, much in the same way as the reworked bioclasts have been preserved. Decomposition of soft parts and the organic parts of the shell by sulphate-reducing bacteria can also change the local redox potential of the pore-water surrounding the shell (Canfield & Raiswell, 1991b), lowering pH and Eh levels and mobilising dissolved Fe^{2+}, which explains the association of calcitised shells with sedimentary pyrite infills (Fig. 8.3g,h).

Other than the corrosive dissolution of carbonates by microbial decay, there is very little evidence for biological break-down of shells within the indigenous invertebrate fauna. Even the most well-preserved shells show no evidence of having being bored by micro-organisms (Figs. 8.3b and 8.4e), encrusted or drilled by larger predators (e.g. by







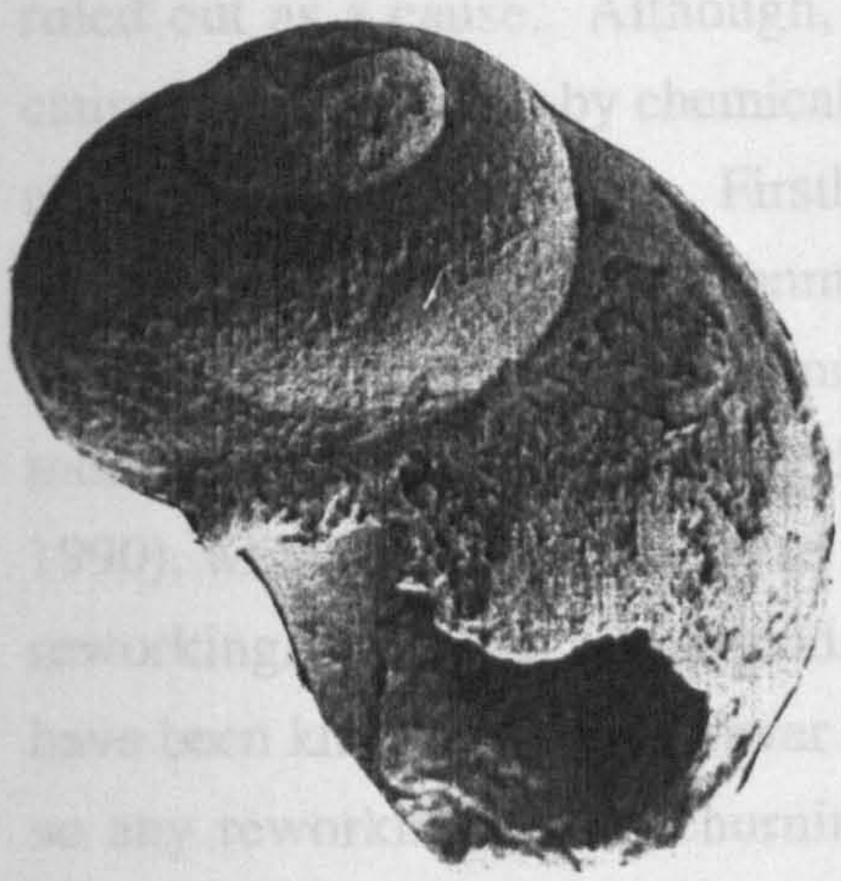


Figure 8.4. More indigenous viviparid gastropod material : (a-d) preserved in ferric iron oxides, (a) G. 89592, squashed gastropod (10.8mm width); (b) G. 89595, gastropod fragment (9.3mm width); (c) G. 89597, squashed gastropod (7.2mm width); (d) G. 54033, broken gastropod; (e) G. 56633, showing scratches on surface (10mm = 250μm); (f-g) abraded calcitised gastropods, (f) G. 51501 (2.7mm width), and (g) G. 51246 (1.6mm width).

other gastropods). However, in modern-day situations drilling is largely restricted to marine species (Brett, 1990). It is difficult to say whether the surface abrasion seen on some of the specimens (Figs. 8.3c,d,g and 8.4d,f,g) may have occurred upon surfaces structurally weakened by bioerosion. Larger predators such as the semionotid fish *Lepidotes*, may have been responsible for the crushing and fragmentation of some of the larger shells (e.g. Figs. 8.3f and 8.4b), but this is extremely difficult to elucidate because of their poor preservation as casts. Small-scale parallel scratches have been seen upon a number of the well-preserved fossils (Fig. 8.3e), but their formation remains elusive. It is most likely that the breakages and scratches have been produced by mechanical processes such as bioturbation and pedogenic mixing.

Many of the carbonate shells show mild surface crumbling or dulling (Figs. 8.3c,f,g and 8.4d,f,g). Brett & Baird (1986) have suggested that this can be attributed to three different causes: biological corrosion or bioerosion, geochemical dissolution of shell carbonate as a result of chemical instability within sediment pore-waters and finally, physical grinding and polishing or abrasion. They also suggest that in cases were the causes of the surface erosion are not easily resolved, then the term "corrasion" should be applied (Brett & Baird, 1986). In the case of the Hornsleasow fossils, bioerosion can be ruled out as a cause. Although, it is more difficult to state whether the erosion was caused by abrasion or by chemical dissolution, a few factors seem to point to the latter as a more probable candidate. Firstly, the gastropods appear to be *in situ* within the pond, which was a fairly quiet environment where rapid burial took place. Secondly, the shells are quite tiny (most less than 5mm) and the majority are unbroken. Small bioclasts are much more likely to be damaged and fragmented by abrasion than larger ones (Brett, 1990), which suggests that these have not been exposed to a great deal of transport and reworking. However, gastropods have fairly dense and compact univalved shells and have been known to survive over a thousand hours of continuous tumbling (Brett, 1990), so any reworking by the churning effects of bioturbation and soil-forming processes would have produced only minor physical damage. These processes may have led to the fragmentation of many of the larger shells which are preserved as ferruginous nodules (e.g. Fig. 8.3a-c), but this mechanical reworking could have taken place after permineralisation. It is more likely that the corrasion of the carbonate bioclasts was caused by slight dissolution of the replacement calcite. Carbonate shells are commonly etched within brackish and fresh-water settings (Alexandersson, 1972), and especially within marsh or boggy conditions where chemical instability of the carbonate in the slightly acidic pore-waters causes surficial leaching and dissolution. Bioturbation and degradation of organic material commonly increases the amount of dissolution by introducing fresh influxes of weak organic acids (Brett, 1990), and this is particularly true where shells have become concentrated in carbonate-saturated sediments.

8.2.4. Time averaging

The accumulation of invertebrate fossils in the Hornsleasow clays, does not represent an *in situ* living community, but a death asssemblage of mixed provenance (a 'mixed' assemblage *sensu* Kidwell & Bosence, 1991), including indigenous freshwater and derived or reworked marine forms. It is also likely that this assemblage has been 'time-averaged'. In crude terms the presence of reworked material from the underlying limestones ('remanie' assemblages: Kidwell & Bosence, 1991) and the 'piped' fauna introduced into the hollow from the overlying and younger Sharps Hill Formation (Ashford, 1992) are evidence for gross-scale time averaging. However, it is also probable that the *in situ* elements, the gastropods, *Valvata* and *Viviparus*, and plant matter, along with marine invertebrates derived from contemporary storm washover (foraminiferans and ostracods), were mixed after death within the pond and during burial in the clay units.

In life, freshwater viviparids are epifaunal gastropods, and must therefore be buried rapidly to be preserved in life position. Although, the sieving tank yields only crude fractions of residue, it was possible to find shells of the same preservational mode within the sieved fractions of both paleosol units. Although the shells within these accumulations probably underwent the same taphonomic history after death, it is more difficult to prove that they were associated in life. Similar accumulations of shells were carefully excavated on site, and were shown to have random orientations within the clays. Most shells also exhibit some sort of post-mortem breakage, fragmentation or abrasion. This could have occurred before or after burial, but indicates that the shells underwent a degree of transport. Physical damage could occur in the pond by limited current action and some may be attributed to predator action, but most would have been incurred by animal (mainly vertebrate) activity stirring up of the basal muds of the pond. In modern-day crocodile or alligator ponds, the animals are known to maintain and enlarge their environment by excavating the sides and base, particularly in areas which suffer seasonal drought (N. Rutger, pers. comm., 1992). Large scale bioturbation at the water-sediment interface, would be accompanied by small scale invertebrate bioturbation and pedogenetic churning of the buried clays. As sedimentation rates in Grey Clay times appear to have been fairly low, bioturbation and soil movement were the main forces serving to thoroughly mix buried fossil remains with fresh dead material. Following silting of the pond environment in Green Clay times, the fossil soil horizons would continue to be mixed on a slower-scale by pedogenetic and bioturbation effects, and any derived marine invertebrate material would be incorporated into Green Clay fauna.

Time-averaging of the indigenous fossil remains seems to be on the same scale as the time it took for Grey and Green Clay deposition, followed by their horizonation and maturation into fossil soil units. Depositional time cannot be estimated as sedimentary

textures have been homogenised by the subsequent pedogenesis. However, an estimation of the time taken for the soils to mature can be made (section 5.12). The Grey Clay represents a fairly immature to moderately developed fen-type paleosol, forming over a period of 100 - 10^4 years (section 5.12.1). The seasonally-wet Green Clay brakeland paleosol is much more mature and could have taken anything from 10^4 - 10^5 years to form (section 5.12.2).

In conclusion, the low-density faunal assemblage within a large homogeneous sedimentary body suggests that the Hornsleasow fossil invertebrate concentration represents a "within-habitat time-averaged assemblage" (*sensu* Kidwell & Bosence, 1991). This is supported by the fine-scale heterogenity in composition which has been caused by incomplete reworking (Fürsich, 1978) and the apparent mixture of shells showing differing degrees of physical, chemical and biological alteration derived from different palaeocommunities. This concept of time-averaging in the fossil soils, also applies to the vertebrate and plant remains.

8.3. The Hornsleasow paleosol floral assemblage

8.3.1. The palynofloral assemblage (Table 8.3)

The palynofloral assemblage from eight untreated clay samples and two from the overlying carbonate sands were studied by J. Cole (1989) of Gearhart Geodata Ltd., Aberdeen. The clay samples included three from both the Green and Grey Clays, and two samples from the plant-rich surface 'A' horizon, termed by Cole, the 'mixing zone' (Fig. 8.5). The majority of the palynomorphs recovered from the clay samples are of terrestrial origin, whilst those in the overlying 'transgressive' sands are a mixture of marine and non-marine forms (Table 8.3).

By far the highest palynomorph recovery was from the Grey Clays, especially samples 1007 from grid squares 5E and 3C, although the sample from quadrate 4D was rather impoverished (Fig. 8.5). Within the Grey Clay there are high bisaccate pollen : pteridophyte spore ratios (Fig. 8.5). Bisaccate pollen (Fig. 8.6o,p) is considered to be wind-dispersed, and to originate from upland conifers, and is therefore derived from the nearest land areas (e.g. the London-Brabant Massif and Welsh-Pennine landmass) or from local dryer highs upon the palaeokarst. Cole (1989) suggested that the pollen may have been derived from areas over 10 kilometres distant from the pond and that the high ratios indicate low sedimentation rates for the Grey Clay, which supports the sedimentary evidence for periodic flooding deposition, rather than a fluviatile regime. However, Bathonian conifers are known to have inhabited low-lying environments much more than at present (for example, the Scalby Formation floodplains of Yorkshire; Boulter & Windle, 1993) and therefore, a local derivation for some of the pollen cannot be ruled out (J. Cole pers. comm., 1992). For instance, in the Grey Clay spores of the coniferous

UNIT	BIOTIC ASSEMBLAGE	ECLGY	TAPHONOMIC ASSEMBLAGE	DERIVATION (in clays)	REFERENCE	FIG. 8.5 LEGEND
	PALYNOFLORA					
Gy	*Botryococcus braunii* (green alga)	A(fw-olg)	well preserved	*in situ*-pond	C-89, M-92	A
Tr	Chorate dinocysts (red alga)	M	well preserved	n/a	C-89	B
Gy/Tr	*Osmundacidites [Verrucosisporites]* sp. (fern)	T-bog	rare	local flora	C-89	C
Gy (Gn) Tr	*Microverrucosisporites* sp. (fern)	T-bog	rare	local flora	C-89	D
Gy/Tr	*Ischyosporites [Klukisporites]* sp. (fern)	T-bog	rare	local flora	C-89	E
Gn (Tr)	*Laevigatosporites* cf. *couperi*	T-bog	rare	local flora	C-89	F
Gy	(fern/lycopod)	T-SA	very rare	local flora - *in situ*	C-89	G
	Staplinisporites [Microreticulatus] sp. (lycopod)	T-SA	common; well preserved	local flora - *in situ*	C-89, M-92	H
Gy/Gn/Tr	*Densoisporites velatus* (lycopod)	T-upland	rare	derived	C-89	I
Gy (Gn) Tr	*Deltoidospora [Cyathidites]* sp. (gymnosperm)	T-upland	well preserved	local flora-coastal	C-89, M-92	J
Gy (Tr)	*Callialasporites* sp. (conifer)	T-upland	well preserved	local flora-coastal	C-89	J
Gy	*C. trilobatus* (conifer)	T-upland	rare; well preserved	local flora - coastal	C-89	J
Gy	*C. dampieri* (conifer)	T-upland	common; well preserved	windborne	C-89, M-92	K
Gy (Gn)	Bissaccate pollen (coniferous)	T	poorly preserved; rare	?local flora	C-89	L
Gn	Verrucate monolete (undifferentiated)	T	Very rare	?local flora	C-89	M
Gy/Gn	Cingulate spore (undifferentiated)					
	PLANT DEBRIS	T-upland	Lignite/coalified	local flora	M-92	
Gy (Gn)	Wood material (?horsetail/conifer)	T-bog	Perfect - cell structure	local flora	M-p	
Gy (Gn)	Charcoal fragments (lycopod/fern)	T-bog	Iron oxides/coalified	*in situ*	V-89, M-92	
(Gy) Gn	Rootlets (lycopod)					

Table 8.3. Hornsleasow plant biota recovered from the clays and overlying transgressive unit, showing palaeoecology, taphonomy and possible derivation. Abbreviations as for Tables 8.1. and 8.2

SAMPLE	LITHOLOGY	PALYNOMORPH RECOVERY 0 20 40 60 80 100	PALYNOMORPH DIVERSITY 0 5 10 15 20	A	B	C	D	E	F	G	H	I	J	K	L	M	SPORE : BISACCATE RATIO % 0 50 100
4D 1005	Transgression			-	-	-	-	-	-	-	-	-	x	o	-	-	
5E 1009	Transgression			-	x	o	o	x	x	-	●	◐	x	◐	-	-	
4D 1010	Mixing zone			-	-	-	-	-	-	-	o	x	-	o	x	-	
3C 1010	Mixing zone			-	-	-	x	-	-	-	-	-	-	-	-	-	
5E 1006	Green Clay			-	-	-	-	-	-	-	-	-	-	-	-	-	
4E 1006	Green Clay			-	-	-	-	-	-	-	-	-	-	-	-	-	
4D 1006	Green Clay			-	-	-	-	-	o	-	-	x	-	-	-	x	
5E 1007	Grey Clay			-	-	-	x	-	-	-	x	x	x	+	-	-	
4D 1007	Grey Clay			-	-	-	-	-	-	-	-	o	-	-	-	-	
3C 1007	Grey Clay	[403]		x	-	x	◐	x	-	x	◐	o	o	+	-	x	

Bisaccates

Spores

Numbers of Specimens

x = 1 o = 2-4
◐ = 5-9 ● = 10-24
+ = >25 - = Barren

Figure 8.5. The distribution of palynomorphs within the clay lens at Hornsleasow, after Cole (1989).

genus *Calliasporites* (Fig. 8.6j, k) are fairly common and these are commonly associated with a lower coastal plain environment (Table 8.3). Conifers are known to be more abundant than other Jurassic gymnosperms (cycads and gingkos) in most British Bathonian palynofloral assemblages (Boulter & Windle, 1993) and this is borne out by the relative rarity of these latter taxa at Hornsleasow (Fig. 8.5). Boulter & Windle (1993) have suggested that this indicates an increased aridity within the Bathonian.

The dominant spore in the Hornsleasow assemblage, *Densoisporites velatus*, (Fig. 8.6k,l) is considered to be lycopodaceous (Dettman, 1963; Table 8.3). The spores are probably autochthonous, and the clubmosses would have formed the common herbaceous flora on the limestone substrate surrounding the deposit (Metcalf *et al.*, 1992). Fern spores are not abundant within the paleosols (Figs. 8.5 & 8.6g,h,i; Table 8.3) indicating a relative rarity of the parent plants surrounding the pool in comparison to the dominant clubmoss flora. This is in direct contrast to the palynofloral studies upon Middle Bathonian assemblages made by Boulter & Windle (1993), which show a definite 'fern spike' in coastal deposits from Oxfordshire and Northamptonshire and in alluvial Yorkshire sediments. This suggests that the Hornsleasow pond and surrounding karstic floodplain were wetter than these other environments.

The green alga *Botryococcus braunii* is also found within the Grey Clay (Table 8.4, Figs. 8.5 &. 8.6a,b,c). This alga has a long geological record from the Precambrian to the present, and in present-day circumstances is dominant in low-nutrient (oligotrophic) lacustrine environments (Hutchinson 1957). This supports the sedimentary and faunal evidence that the Grey Clay was deposited within clear standing water, with little stirring up by current action. The virtual absence of dispersed kerogen in the palynofloral slides (Fig. 8.6a) also supports a low-energy environment with little fragmentation of wood derived from the local flora by fluviatile influence (J. Cole, pers. comm., 1992).

In contrast the Green Clay has yielded only very rare or barren palynomorph assemblages (Fig. 8.5). The assemblage is similar to that within the lower clays. The clubmoss palynoflora is still dominant (Table 8.3) suggesting that the minor change in soil types across the Grey-Green Clay boundary was not a major environmental event affecting the whole of the karstic terrain (Metcalf *et al.*, 1992). However, more importantly the green alga *B. braunii* is not present in the upper clay units, supporting the sedimentary and palaeontological evidence that the pond conditions were transient (Table 8.3).

The 'mixing zone' clay (1010) is the yellow-green mottled 'A' horizon pervaded by many vertical rootlets at the top of the Green Clay paleosol unit (Fig. 5.7). It was also known by the excavating team as the 'plant horizon' (Darlington, 1988). However, the palynomorph recovery and diversity from this horizon was fairly low, although higher than for the rest of the Green Clay (Fig. 8.5). The ratio of wind-blown pollen to spores

is lower than within the Grey Clay (Fig. 8.5), which would suggest much more rapid accumulation of sediments, or that poor taphonomic conditions were responsible for destroying the more delicate pollen grain cuticle.

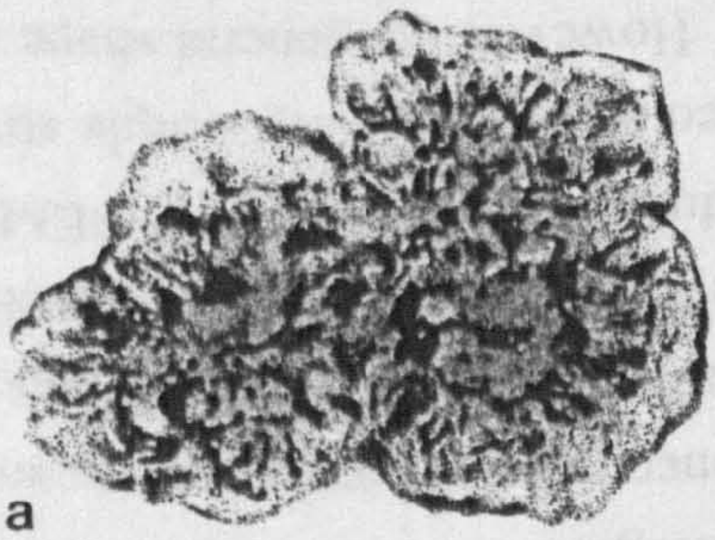
The overlying transgressive sand unit has been described as a beach unit (section 5.13) and consequently they contain marine palynomorphs, such as chorate dinocysts (Fig. 8.6d) and microforaminiferal cuticles (Fig. 8.6e), as well as the terrestrial spores (Table 8.3). The palynomorph recovery is quite high for the transgressive sands and preservation is also fairly good. These factors and the low bisaccate:spore ratios suggest that the carbonate sands were deposited fairly rapidly over the karstic landscape.

8.3.2. Large plant remains

Megafloral remains are confined to the clay units at Hornsleasow, although elsewhere pieces of coalified wood or 'lignite' have been reported within the other facies of the Chipping Norton Formation (pers. obs., 1992). Some of the more sandy facies in eastern Oxfordshire are littered with plant material, in the form of larger pieces of lignite and charcoal, and as microscopic 'specks' of inertinite. Both clay units contain much small scale disseminated fragments of carbonised wood, and even some larger fragments of coal, however, the best preserved megafloral remains were recovered from the base of the Grey Clay, overlying the palaeokarstic surface (Vaughan, 1989, Metcalf *et al.*, 1992). These were preserved either as iron-stains upon the limestone or as part lignified-cellulose and part iron-oxide-replaced compressions. They are either lying flat upon the karstic surface or are banked up against the sides of the hollow and show no obvious signs of alignment (section 5.2; Metcalf *et al.*, 1992). The large pieces of wood which are up to 1.3 metres in length, do not appear to be *in situ* tree-stumps, but allochthonous pieces of timber and branches derived from the surrounding floodplain (Metcalf, 1994). The wood is almost certainly derived from a gymnosperm, and probably can be considered to be coniferous, however, the lack of preserved diagnostic anatomical detail means that identification beyond this is impossible.

Some of the smaller wood fragments and coalified compressions were studied under the SEM to facilitate diagnosis of taxonomic and taphonomic characteristics. These specimens were carefully cleaned by gentle washing with distilled water or agitation within an ultrasound bath for 3-5 minutes to remove clay particles adhering to the surface. They were then allowed to air dry, before mounting onto metal stubs. Most of the specimens are preserved as lignites and consequently lack any discernible primary structures (Fig. 8.7a). A few specimens exhibit partial or complete charring and as preserved as anatomically structured semifusinite or fusinite (= fossil charcoal; Harris, 1958, Cope, 1980). These are described and their significance reviewed in section 8.3.3. They have been identified as conifer (G. 50242; Fig. 8.7i-k), fern (G. 85590; Fig. 8.8a) and fern or possibly cycadophyte (G. 50926/7; Fig. 5.8b-f) by T.P. Jones

Figure 8.6. Palynomorph assemblage (from Cole, 1989) : (a) *Botryococcas braunii* (x690); (b) *B. braunii* (natural light, x690); (c) *B. braunii* (flouresent, U.V. light, x690); (d) Chorate marine dynocyst from faction 1009, grid square 5E (x 690); (e) Chitinous microforaminiferal lining from 1009, grid 5E (x345); (f) Algal cyst (x690); (g) *Osmundacidites* sp. from 1009, grid 5E (x700); (h) *Microverrucosisporites* sp. from 1007, grid 3C (x700); (i) *Ischyosporites* sp. from 1007, grid 3C (x700); (j) Degraded cingulate spore from 1007, grid 3C (x700); (k) *Denoisporites velatus* from 1007, grid 3C (x500); (l) *D. velatus* from 1007, grid 5E (x700); (m) *Callialasporites trilobatus* from 1007, grid 3C (x690); (n) *C. dampieri* from 1007, grid 3C (x690);(o) Bisaccate pollen from 1007, grid 3C (x490); (p) Bisaccate pollen from 1007, grid 3C (x490).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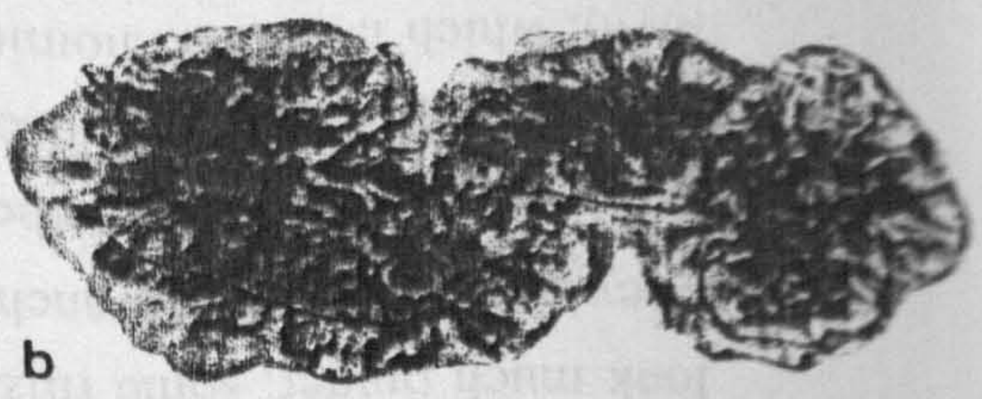
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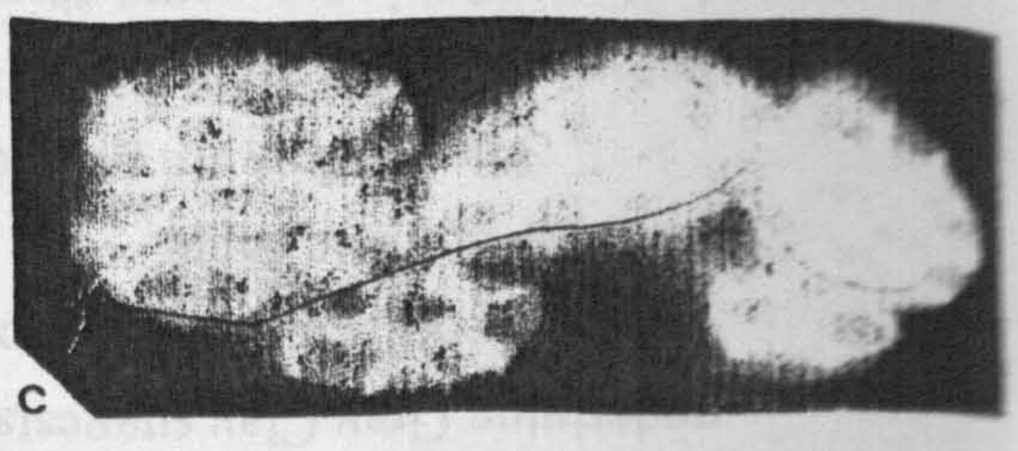
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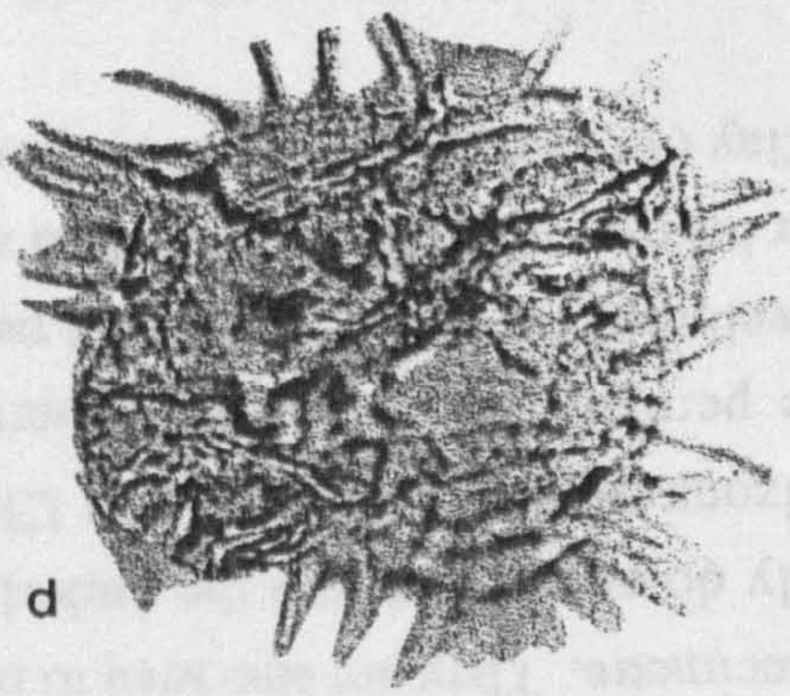
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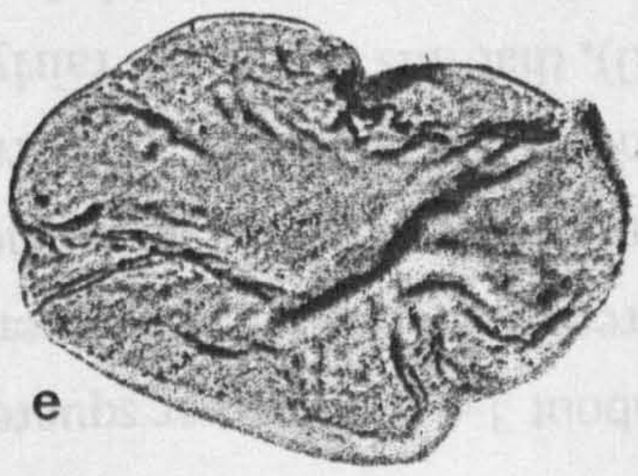
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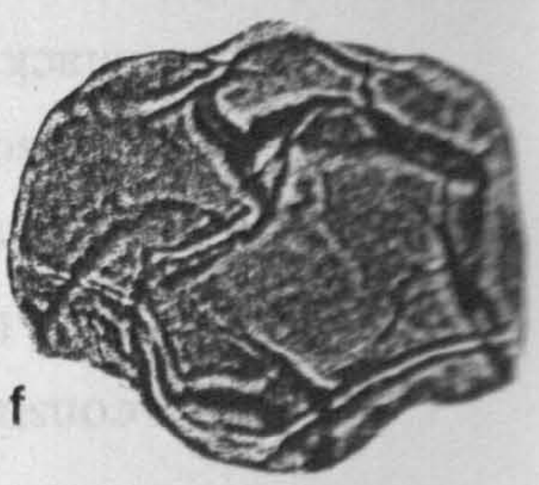
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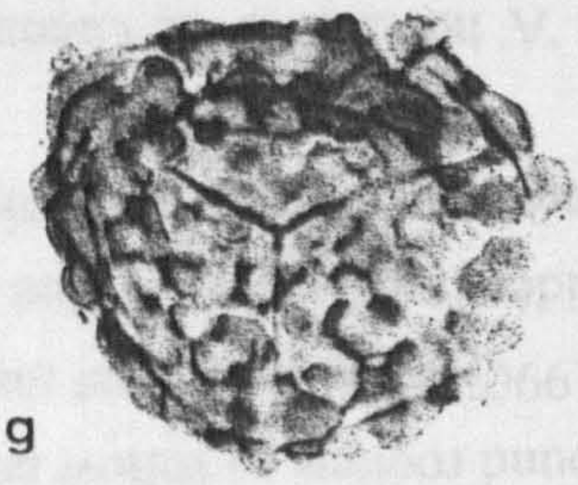
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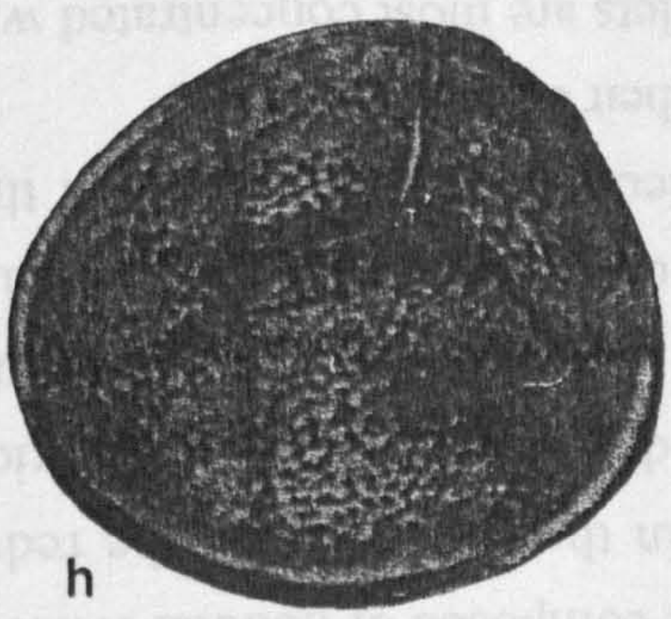
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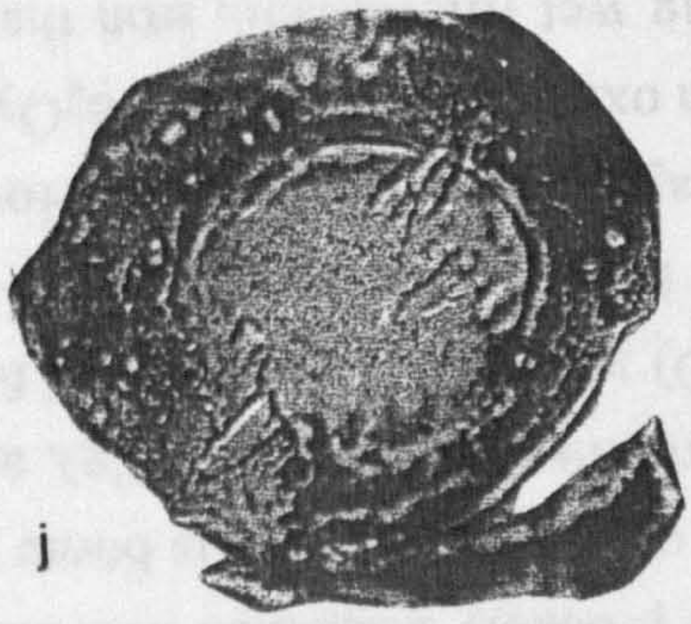
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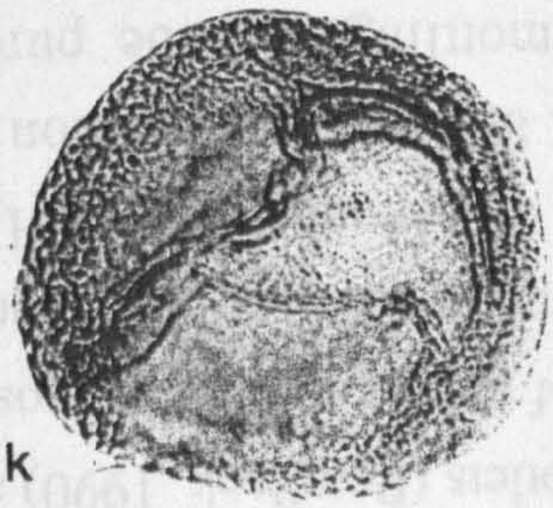
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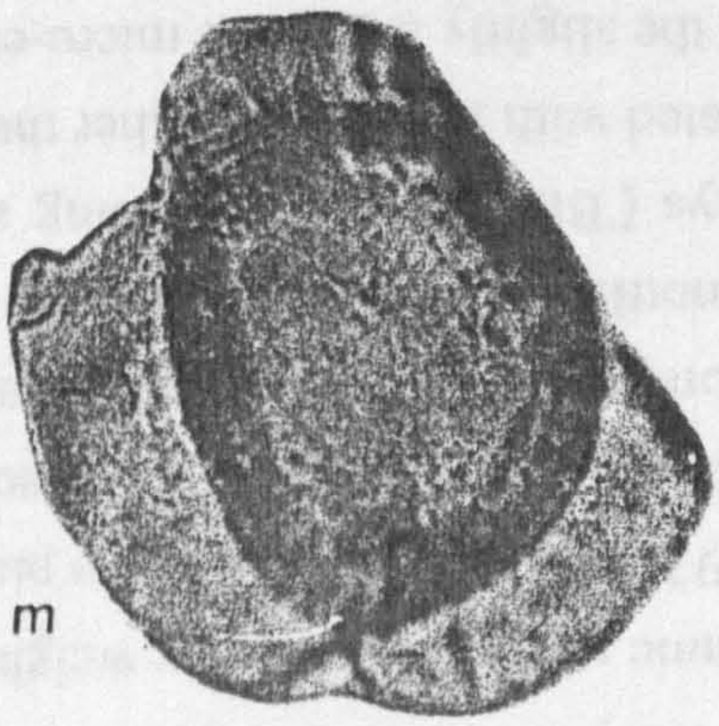
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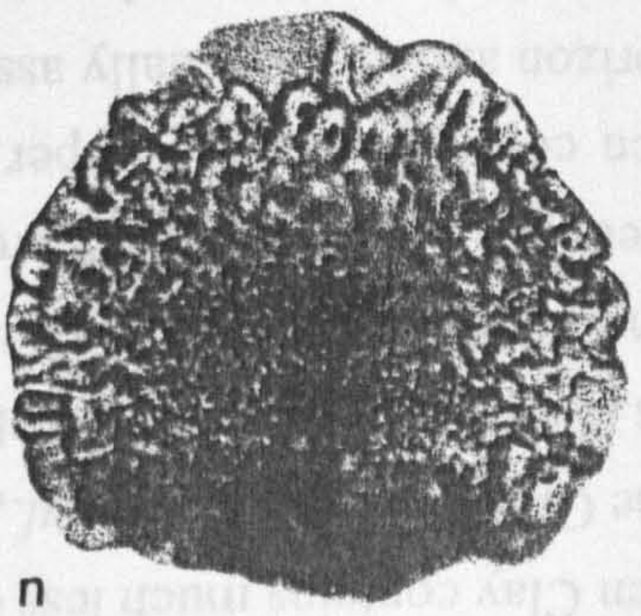
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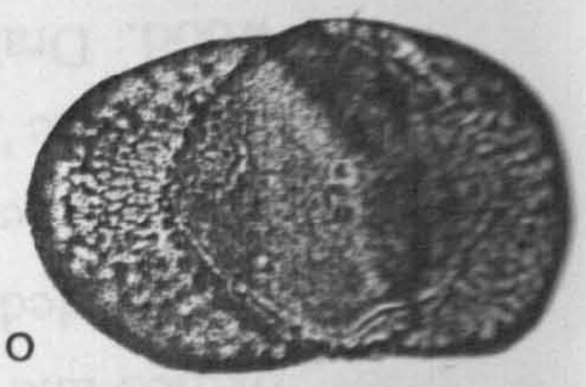
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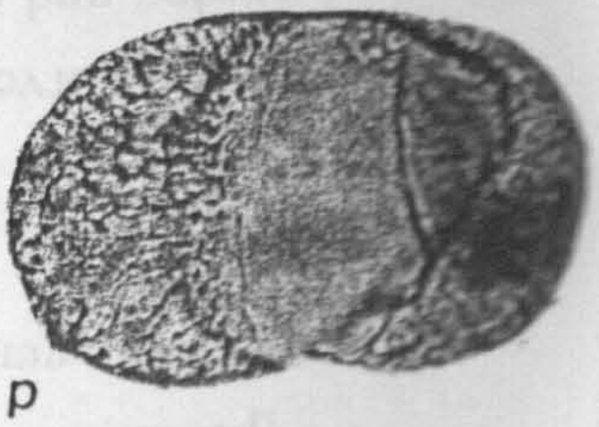
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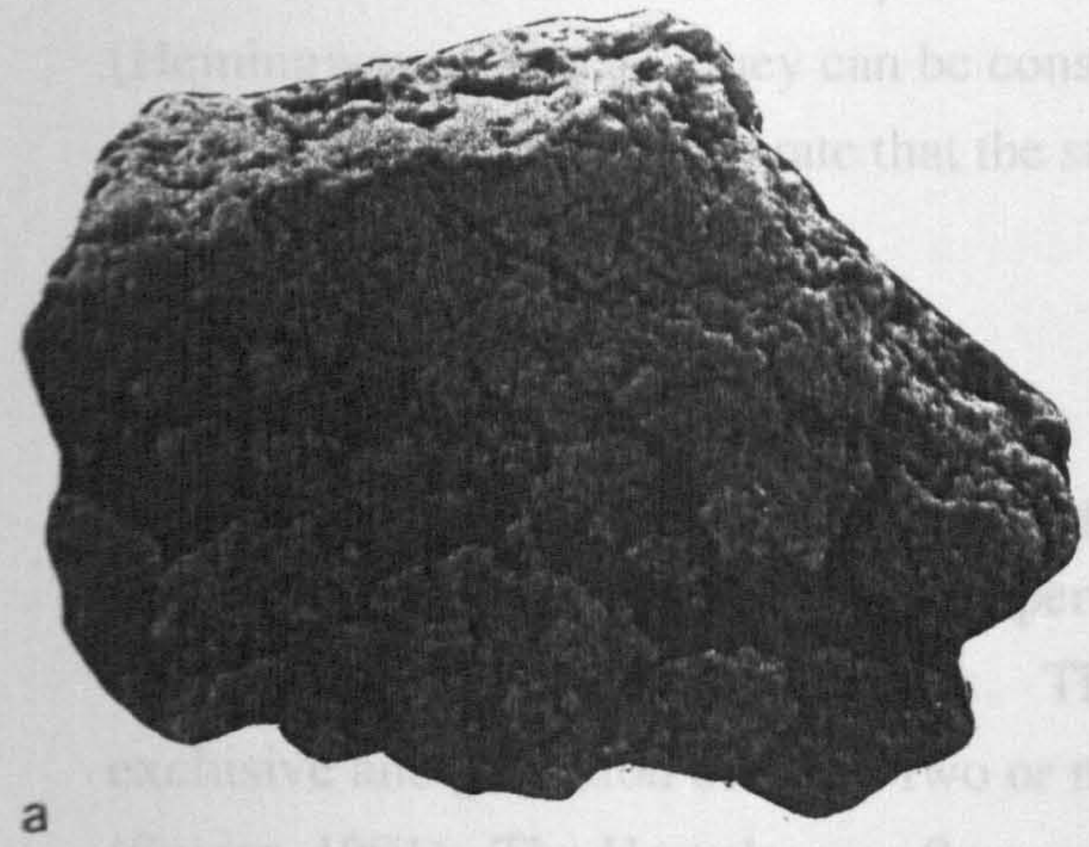
(pers. comm., 1994) and M.J. Cope (pers. comm., 1992). These identifications fit in with the general palynofloral assemblage (Table 8.3).

The Green Clay contains much less organic material (0.85% by weight compared with 1.65% in the Grey Clay; Metcalf *et al.*, 1992), and the majority of the plant remains are preserved as carbonised films within the sediments or have been replaced by iron oxides and oxyhydroxides. The carbonised compression fossils in both clay units are surrounded by reduced drab-haloes and grey mottling, which is particularly noticeable against the green coloration of the upper clays ('Bt' horizon). Mottling also occurs within the 'A' horizon and this is usually associated with root traces, rather than pieces of wood. Drab-haloes are formed around roots by the slightly anaerobic micro-environment surrounding the rootlets (Retallack, 1990) and are probably associated with the reduction and mobilisation of iron within the rhizosphere or root zone. This is borne out by the preservation of some of the rootlets either as carbonised traces (Fig. 8.7g), as pyritised forms (Fig. 8.7h), as limonite ($FeO[OH].nH_2O$) stains or in the form of ferruginous rhizocretions (Fig. 8.7f). The formation of the limonitised root traces and associated oxidised yellow mottling may be purely diagenetic after pyrite. However the rhizocretions are composed of nodular ferric iron oxides (e.g. hematite Fe_2O_3) and were formed *in situ* in the soil. Within the reducing wet rhizosphere, iron that has been mobilised in the drab ferrous state can be oxidised around rootlets to yellow and red ferric oxides, and thus form these concretions (Retallack, 1990). It is likely that many of these roots would have become so encrusted with iron oxides (e.g. Fig. 8.7f) that they would have died and decayed. Indeed, some of the rhizocretions contain carbonised organic material within their core.

The rootlets are most concentrated within the 'A' horizon of the Green Clay, were they constitute about 3-4 rootlets per square decimetre. They are also seen to pervade the whole of the Green Clay layer, but generally do not extend into the underlying Grey Clay. Their confinement to the upper horizons suggest that the Green Clay was an aerobic soil, and as roots rarely penetrate permanently waterlogged parts of soils (Retallack, 1990), that this layer was fairly well drained. That they do not pass into the underlying Grey Clay suggests that this lower horizon remained waterlogged after silting up of the pool and that the top of the Grey Clay ('O' horizon) might corresponded to the palaeo-watertable depth.

The rootlets are generally tubular and irregular in cross-section, with a width of between 2-5 millimetres (Fig. 8.7g: although concretions and replacement can make them look much bigger; some rhizocretions reach a width of 10-15 millimetres; Fig. 8.7f). They tend to taper and branch downwards in the soil. Some of the rhizocretions show a surface striation, whilst others exhibit a nodular appearence. The SEM study of a beautifully preserved pyritised rootlet, revealed an anatomically simple structure (Fig. 8.7h), which is not taxonomically diagnostic. However, the general shape and form of

Figure 8.7. Preservation of plant macrofossils. Lignites (a-b): (a) PCF 741, unstructured carbonised wood (6.3mm in length); PCF 741, structured lignite showing remnant longitudinal rays (10mm = 200μm). Iron oxyhydrate coatings or casts (c-e): (c) G. 93900, cast showing fibrous surface detail (9.2mm in length); (d) G. 51357, cast showing fibrous structure (19.2mm in length); (e) G. 51350, coating (8.3mm in length). (f) G. 89601, ferruginised rhizocretion showing radial septa (45mm in diameter). (g) HQS9-2(1), rootlet preserved in cutinite (brown) and inertinite (opaque) (field of view 2.7mm). (h) G. 51457, cellular structure in tangential section of pyritised rootlet (10mm = 100μm). G. 50242 (i-k): (i) Partially charred fossil conifer wood, charred tissues showing structure (1.5mm in length); (j) Pyrite framboids within crack in charred region (10mm = 10μm); (k) Tracheid pits 't.p.' and homogenised cell walls 'h.c.w.' in charred region (10mm = 10μm).



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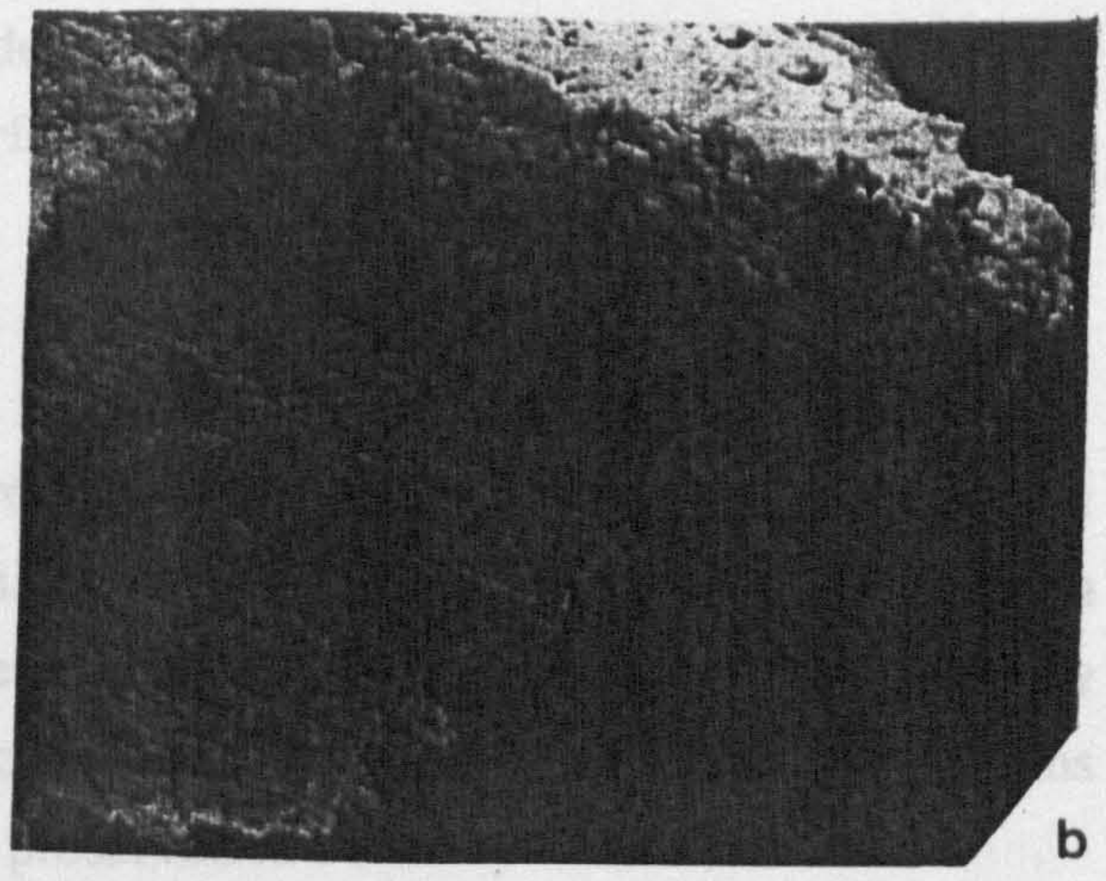

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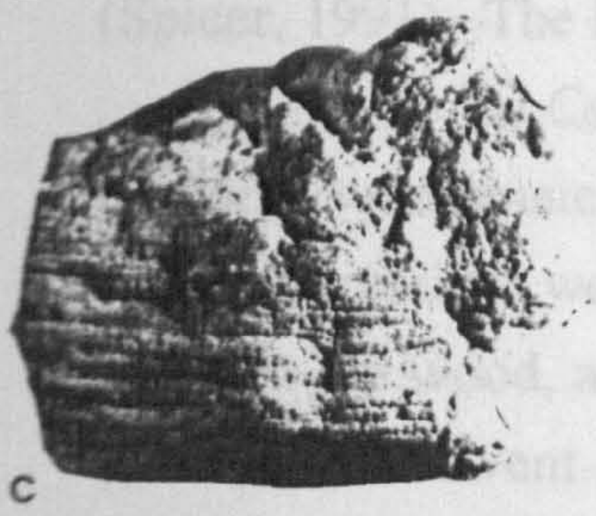

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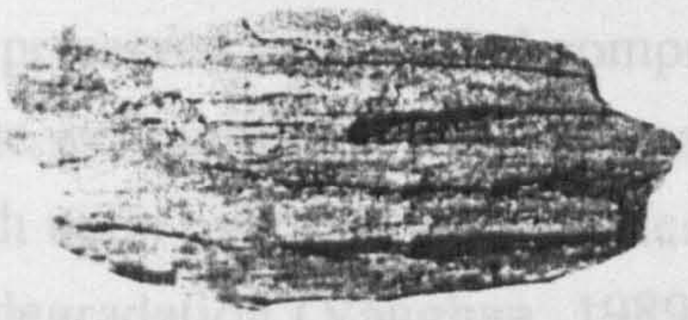

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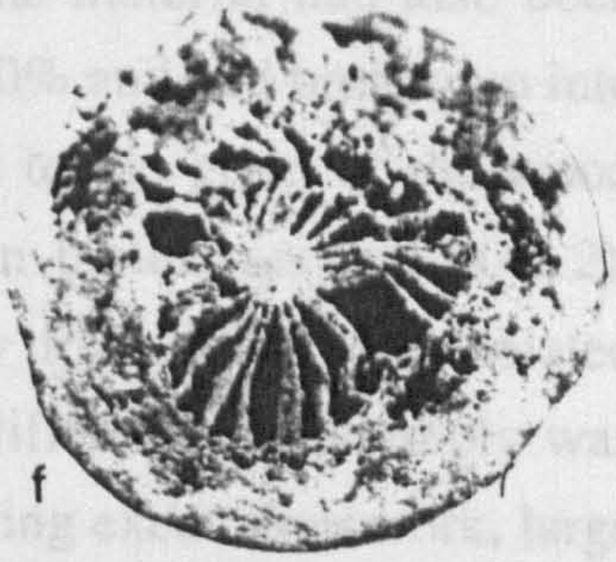

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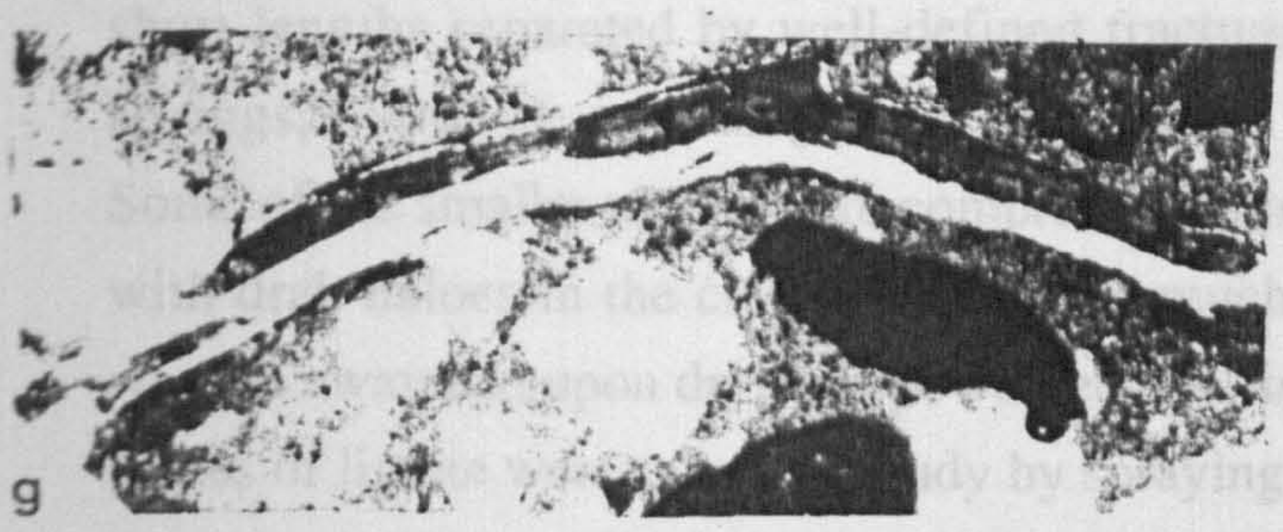

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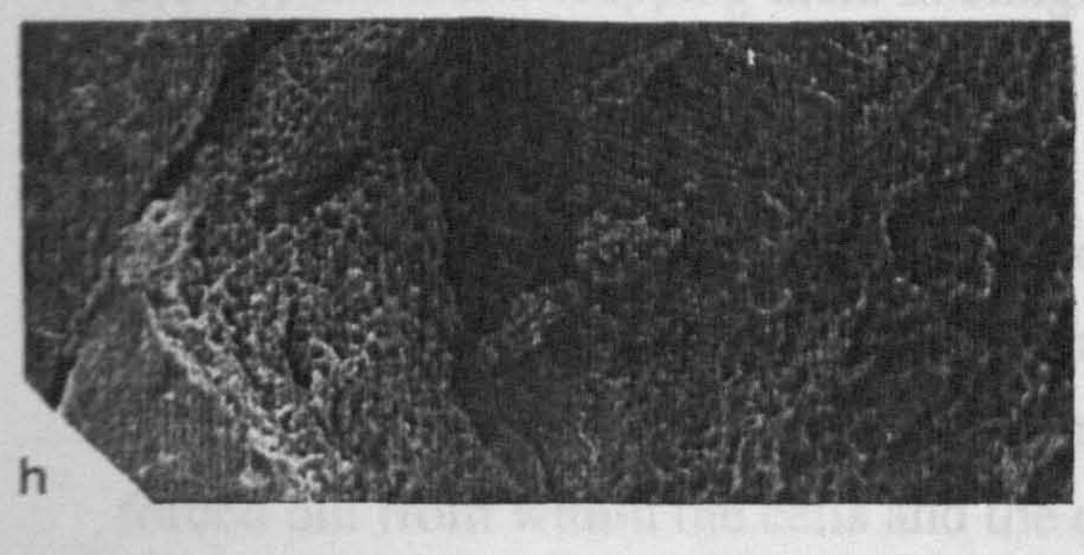

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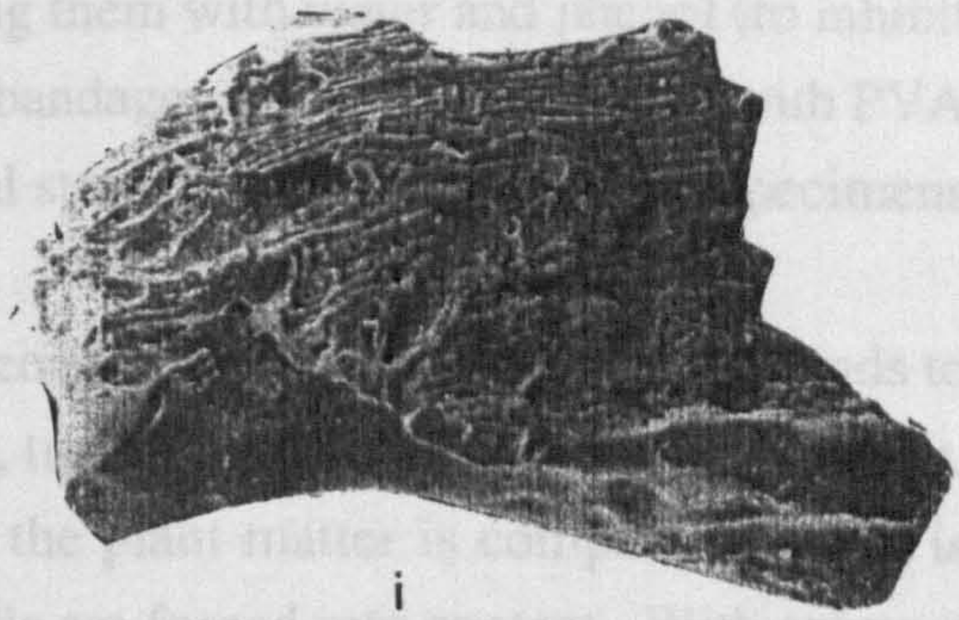

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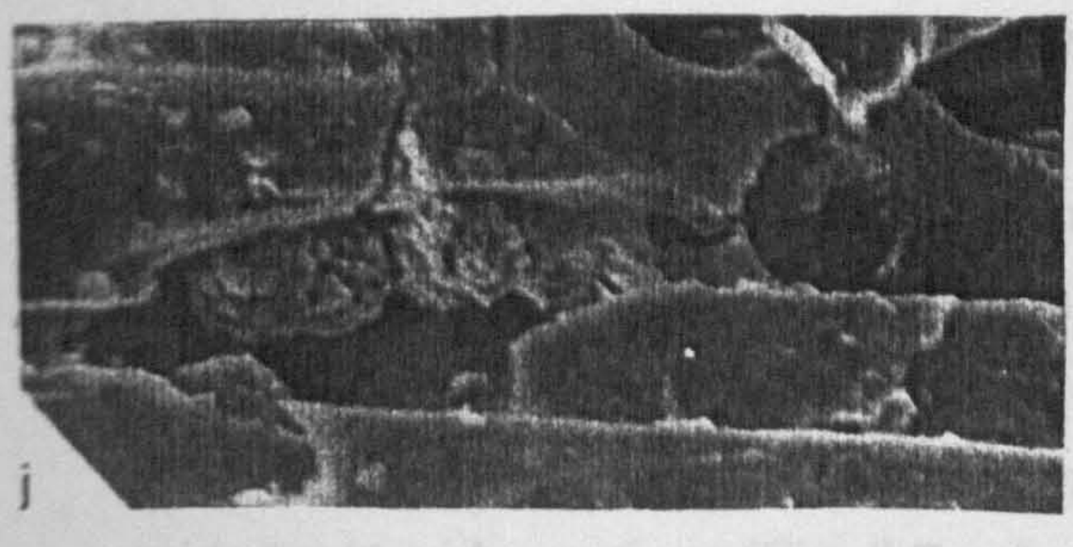

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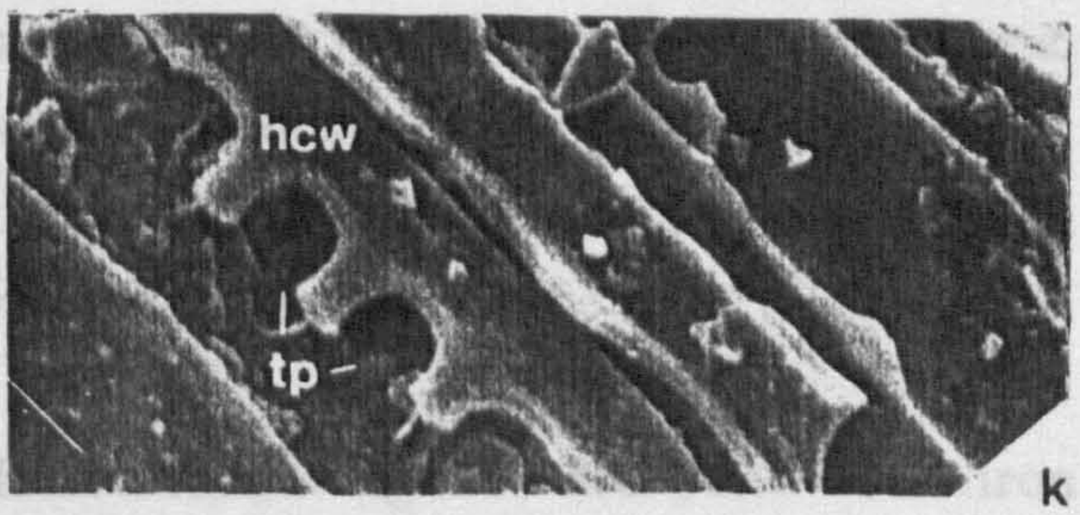

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the rootlets are similar to the *Equisetites* root beds of the Yorkshire Scalby Formation (Hemingway, 1974) and they can be considered to be lycopodacaeous in origin (Metcalf *et al.*, 1992). They demonstrate that the surface of the Green Clay was apparently thickly vegetated.

8.3.3. Preservation of floral remains

Schopf (1975) defined four categories of preservation of fossil floral remains, which are : (a) compressions and impressions; (b) permineralisation and petrifaction; (c) preservation of hard parts (d) casts and moulds. These modes of preservation are not mutually exclusive and gradation between two or more categories can be seen in many specimens (Spicer, 1991). The Hornsleasow flora are preserved in a variety of modes:

(a) *Compressions and impressions - the Hornsleasow lignites.* Most of the large wood material is preserved as coalified compression fossils within the clays. The compressions were quite well preserved showing anatomical textures of the bark and knots in the wood, although upon exposure during excavation, the unoxidised organic material underwent rapid degradation (Vaughan, 1989). The material had also been flattened by the stress of overburden by a factor of about 70-80% and had broken up into short lengths separated by well-defined fractures transverse to the grain of the wood (Briggs, 1990; cf. post-diagenetic 'transverse' fracturing in bones; section 10.7.2). Some of the smaller pieces were compressed into a fine 'coaly' film and were associated with drab-haloes in the clays. On sieving much of this coalified material simply was washed away and upon drying it powdered to a fine dust. During excavation work, large pieces of lignite were saved for study by spraying them with water and phenol (to inhibit decay), and them wrapping them in clingfilm or bandages. Others were treated with PVA solution but, although this preserved the external structure of the wood, those specimens could not be used for microscopic study.

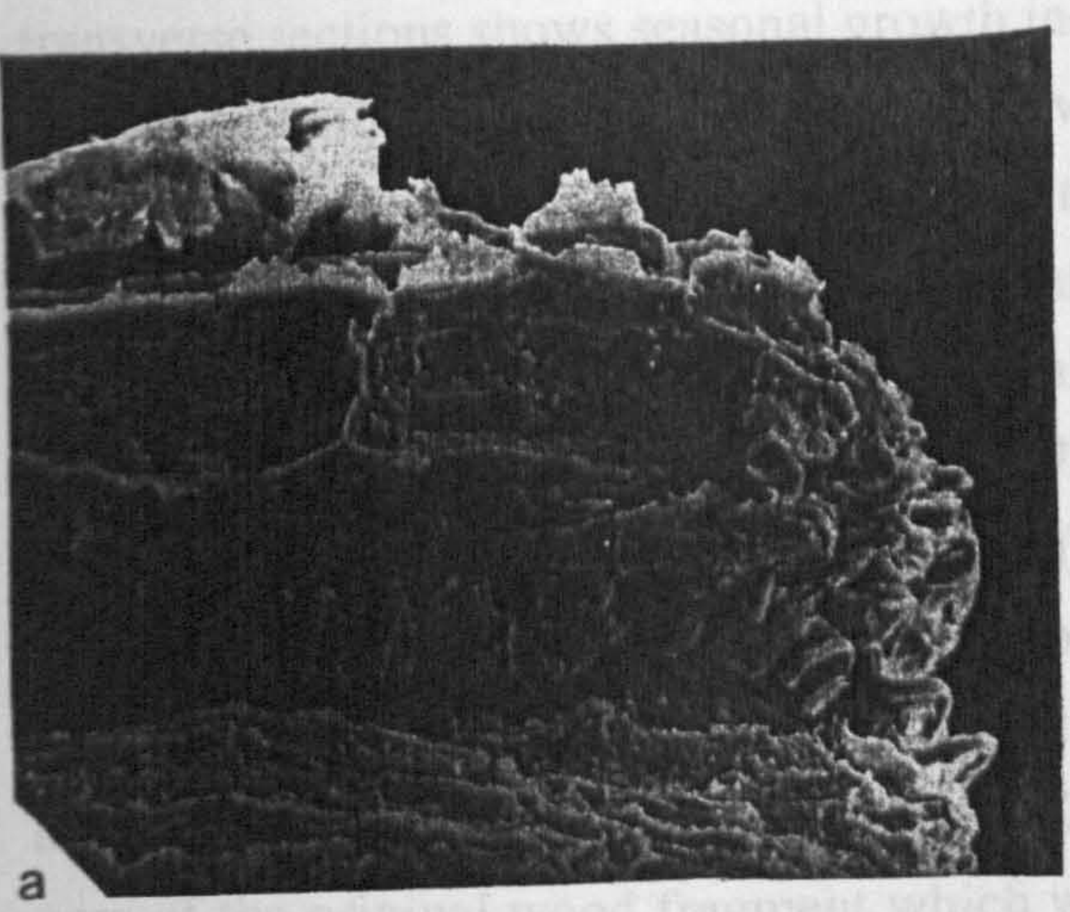
At a microscopic scale compaction and coalification of organic material tends to destroy fine anatomical structures such as cells, tracheid pits and rays (Fig. 8.7a; Scott, 1990a, Jones *et al.*, 1993). This is because as the plant matter is compacted, water is forced out from within the cells and the cell-walls are forced into contact. With extreme compaction the organic matter becomes amorphous as chemical alteration takes place. This process will eventually lead to the loss of other volatiles and the formation of coal. However, the Hornsleasow lignites are only very low rank brown coals and some exhibit the layered remnants of the longitudinal rays when viewed under the SEM (Fig. 8.7b; T. Jones, pers. comm., 1994).

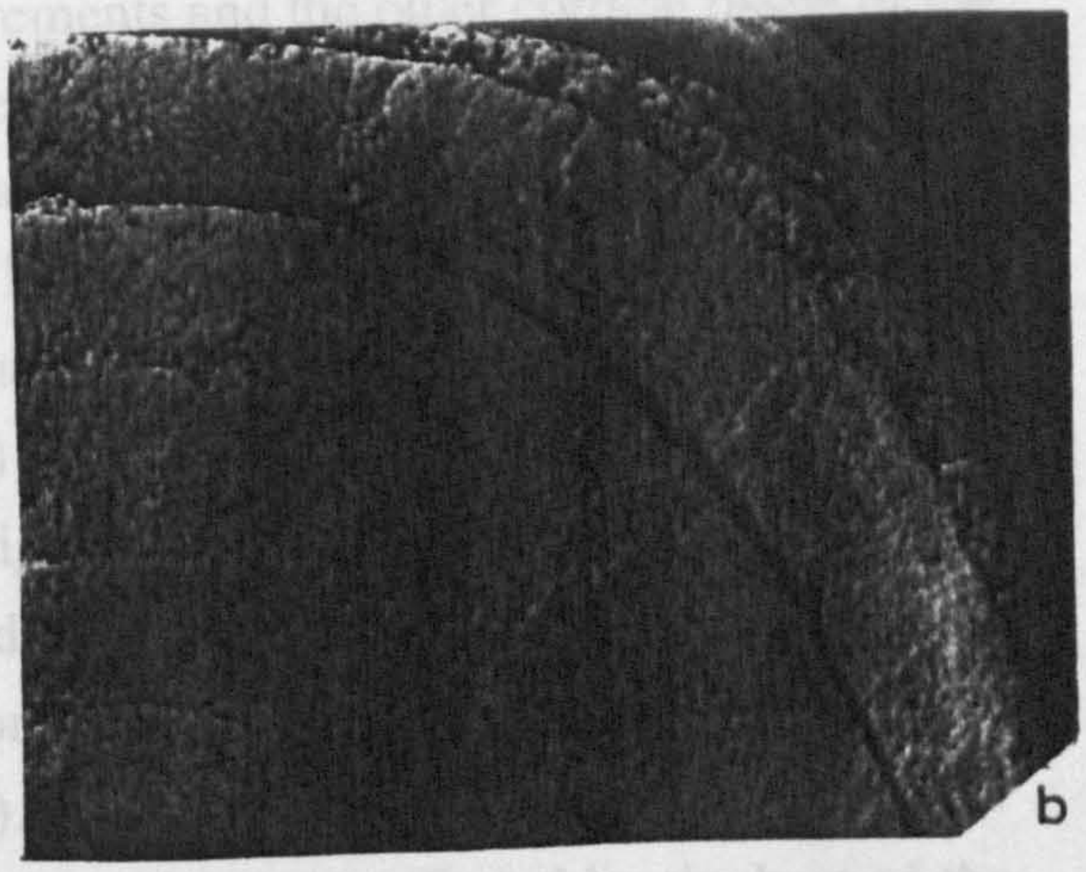
Commonly associated with the Hornsleasow lignites are yellow-red iron oxide/hydroxide coatings (Fig. 8.7c-e). After washing these are the most common plant remains to be found within the sieved residues. They are often finely structured (Fig. 8.7c,d) which suggests that they formed prior to much decay. The most likely cause of

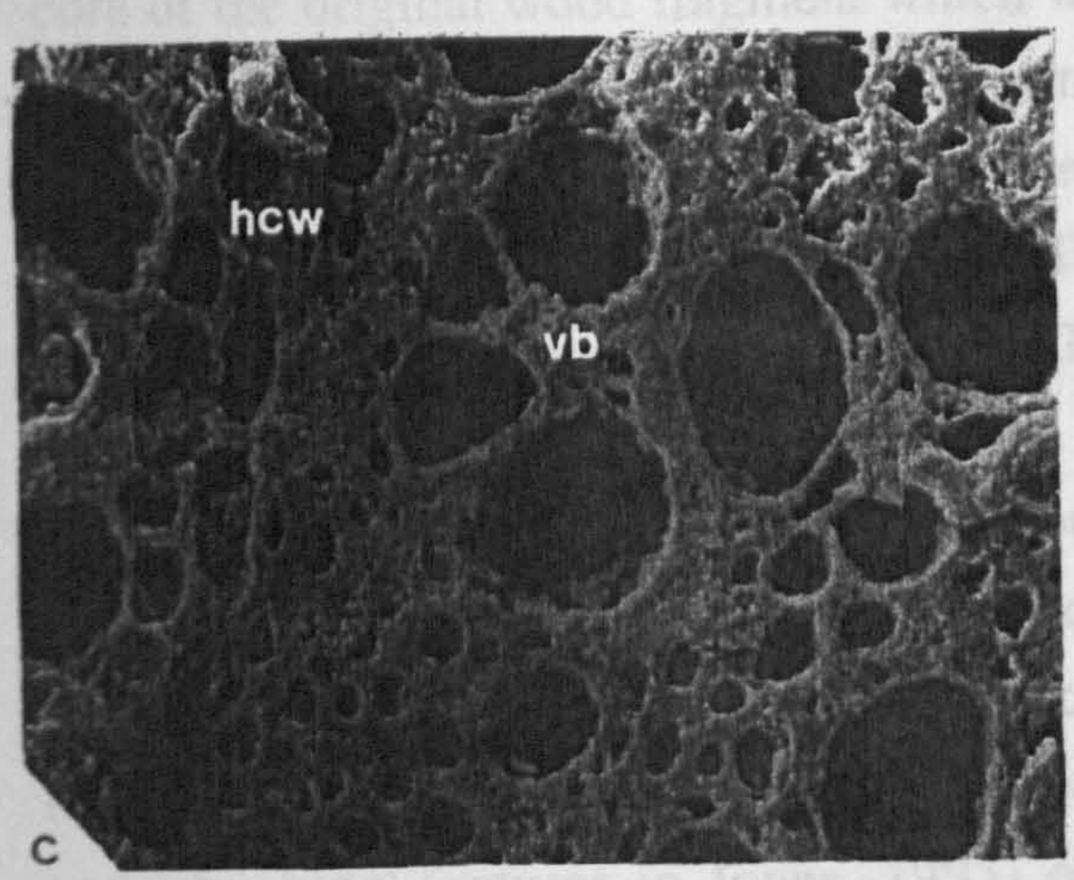
these impression fossils is that they formed prior to burial, as a mineral coating upon the surface of the wood. Iron oxides and oxyhydroxides (e.g. goethite FeO.OH and limonite $FeO[OH].nH_2O$) are known to form a film upon decaying plant matter by bacterial action, wherever there is readily available dissolved iron within fresh-water systems (Spicer, 1991). The coating forms a fine template for impression fossils, and may act to promote permineralisation of the decaying organic tissues after burial. Under the SEM, these casts exhibit some fibrous detail (Fig. 8.7c,d; T. Jones, pers. comm., 1994), although the detailed internal anatomy is not preserved, suggesting decay prior to permineralisation (Scott, 1990a). The coating is also known to inhibit degradation by invertebrate feeders (Spicer, 1991), such as freshwater gastropods, and could be one of the reasons why so much organic material has been preserved. That unoxidised plant material has survived within the paleosol unit, and especially the large pieces of coalified wood within the basal unit, favours anoxic conditions.

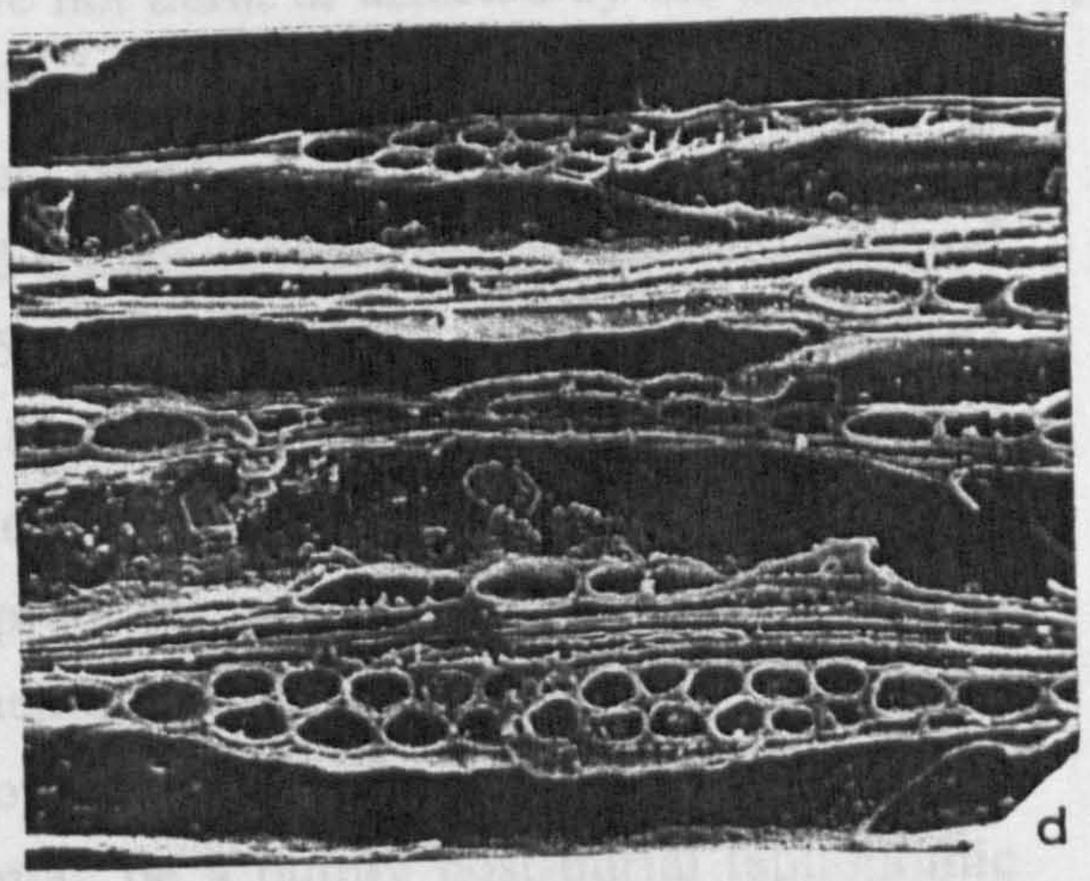
(b) *Permineralisation and petrifaction - pyritisation and the rare Hornsleasow charcoals.* Pyrite is rare in both clay units, but pyrite framboids have been seen within coalified and charcoal fragments under the SEM (Fig. 8.7j). Most pyrite forms as a result of sulphate-reducing bacteria activity, during the first stages of decay (Berner, 1970, 1984). This is often why the less structured, easily degraded of the organic materials are more often preserved as pyrite, whilst the more decay-resistant wood is preserved as lignite (Spicer, 1991). The anoxia required to generate pyrite growth may be extremely localised and even at the scale of cracks and surface depressions within wood material (Fig. 8.7j). However, the complete replacement of organic matter by pyrite is quite rare in the soil: Fig. 8.7h shows detail of a rootlet which has been permineralised by early diagenetic pyrite, preventing tissue compaction. This tiny fragment shows remarkably preserved three-dimensional cellular structure within tangentially broken sections (Fig. 8.7h). It is possible that the yellow-brown limonite stains of wood and rootlets within the clay units represent organic material that was originally pyritised and been subsequently oxidised to iron hydroxides.

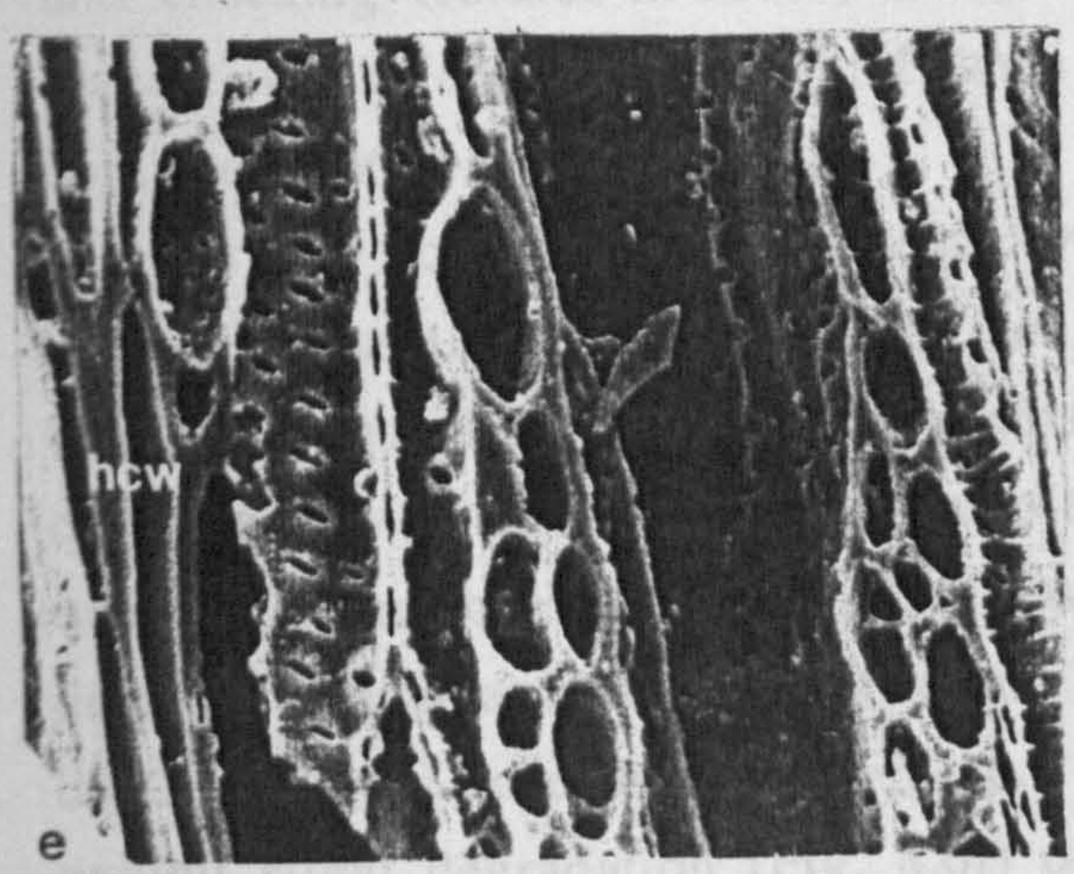
Some of the most exquisite preservation of petrified plant material in the Hornsleasow paleosols is seen in the rare small fragments and flakes of charred wood, which are preserved as fusain or fossil charcoal (Cope, 1980, Scott, 1990a). Much of the fusain within the soils occurs as tiny splinters of 'fusinite' or dispersed kerogen, which could have been deposited within the hollow by windborne or waterborne action from distant forest fires (Cope, 1980). However, large fragments of fossil charcoal also occur, these are typically shorter in length than in cross-section, and typically exhibit fine three-dimensional anatomical detail under the SEM (Figs. 8.7k and 8.8). Specimen G.85590 (Fig. 8.8a) is a small piece of fern charcoal and shows spiralia within the rays. Specimen G. 50926/7 is a small piece of extremely well-preserved fusain (Figs. 8.8b-f). Its is probably cycadophyte or fern in origin (M. Cope, pers. comm., 1994) and in











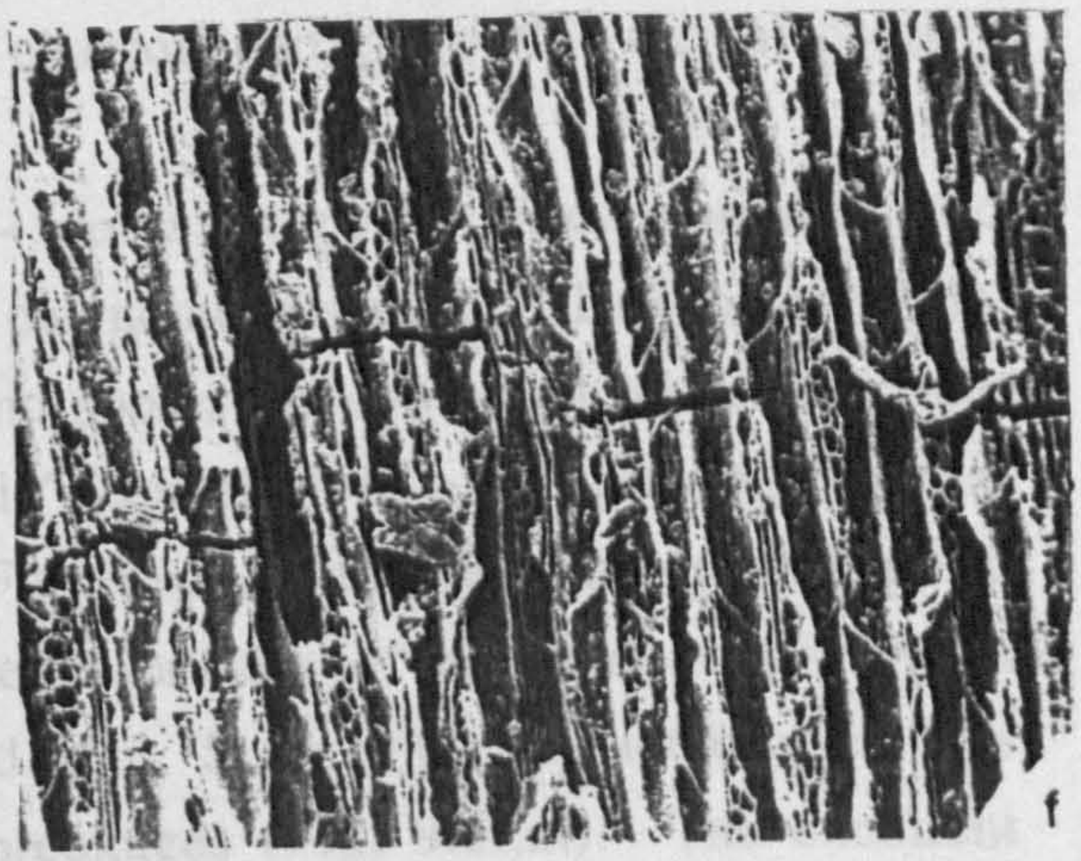


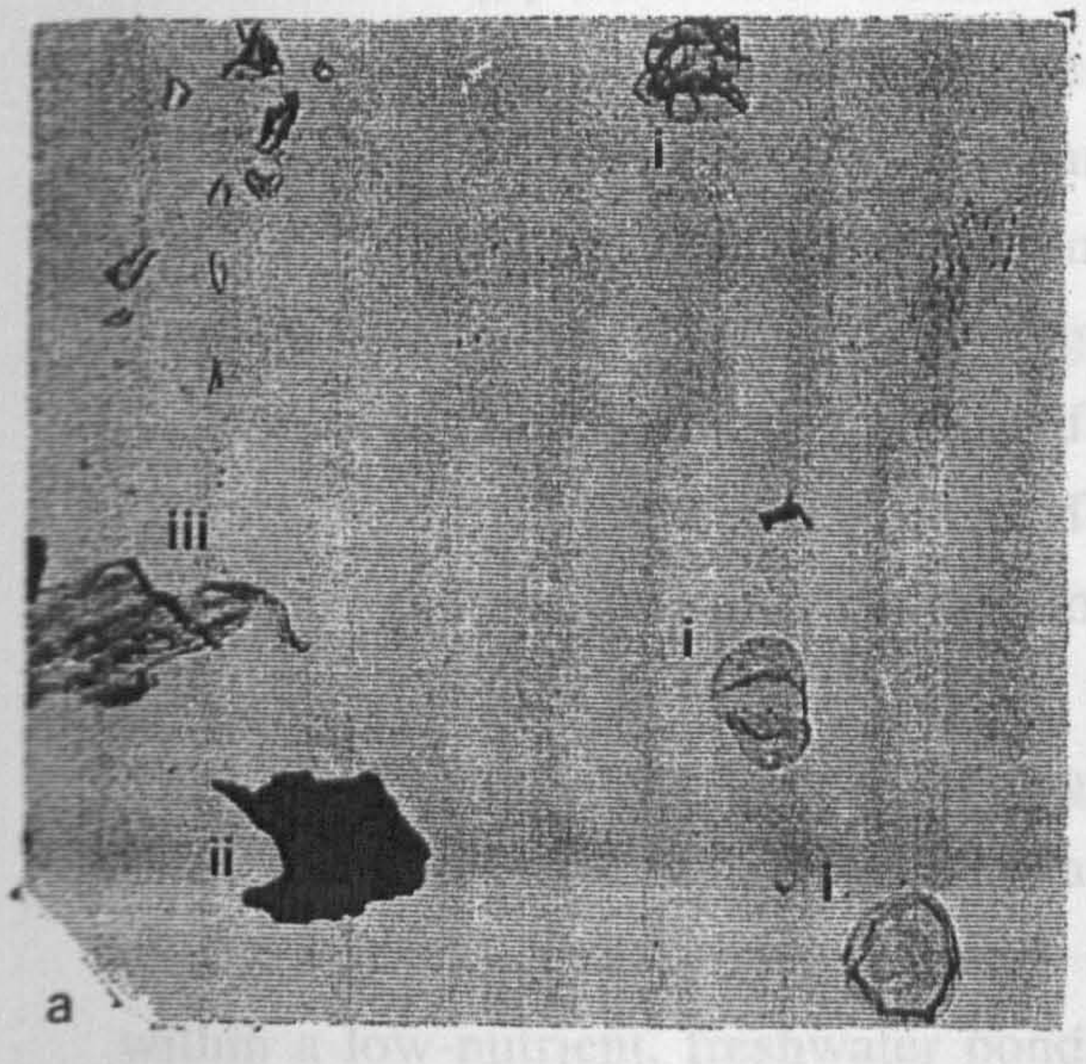
Figure 8.8. Preservation of fossil charcoal. (a) G. 85590, fern fusain in tangential section with spiralia (10mm = 25µm). G. 50926/7, fern or cycadophyte fusain : (b) transverse section showing seasonal growth rings and outer cortical tissue (10mm = 600µm); (c) transverse section showing open lumina 'o.l.' of longitudinal vessels and possible vascular bundle 'v.b.', with homogenised cell walls 'h.c.w.' (10mm = 30µm); (d) tangential section showing tall longitudinal rays comprising 23 cells (10mm = 30µm); (e) tangential section showing 'checking' of homogenised cell walls 'h.c.w.' (10mm = 15µm); (f) tangential section showing abrasion and cracking of fusain (10mm = 75µm).

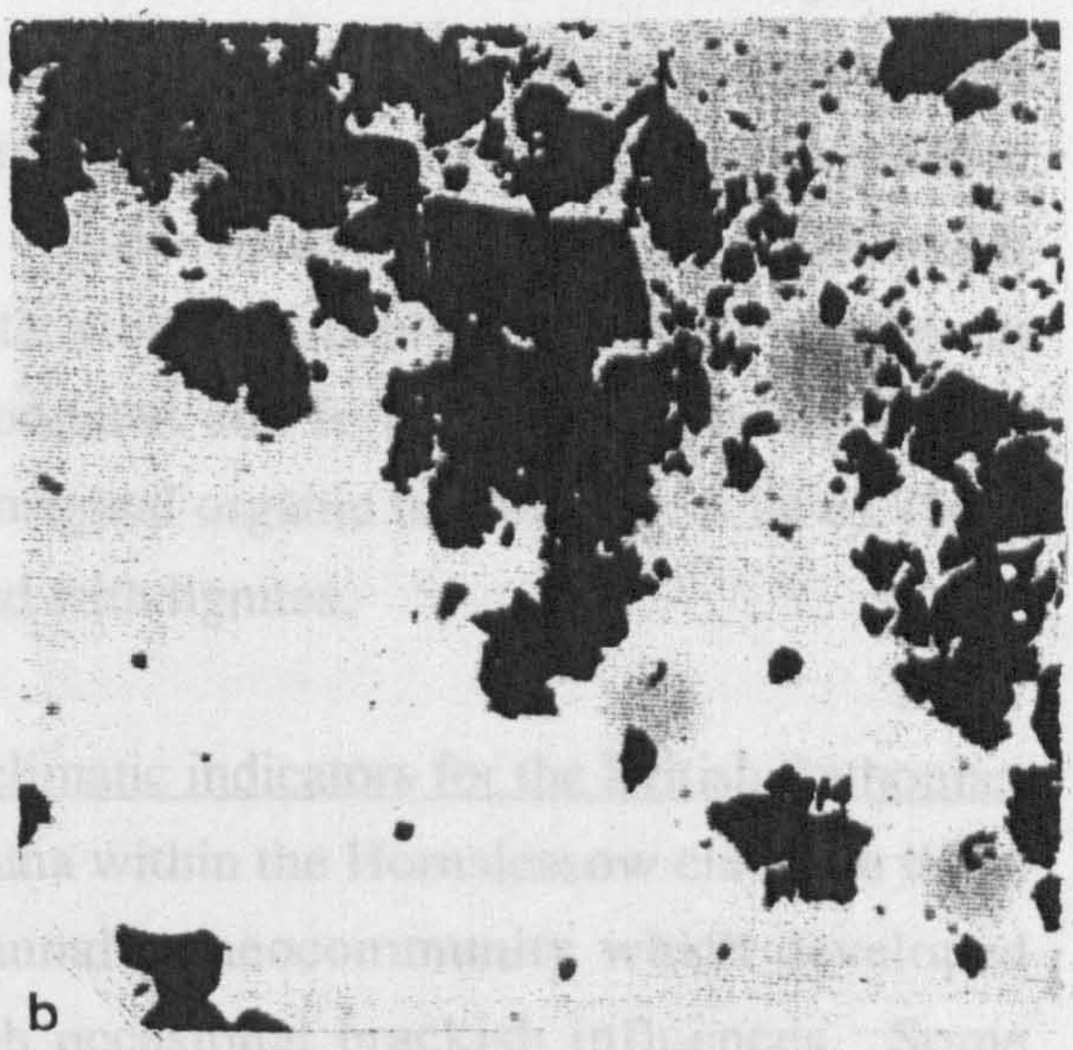
transverse sections shows seasonal growth increments and the outer cortical tissue of the stem (Fig. 8.8b). A close up in transverse view (Fig. 8.8c), shows the open lumina of longitudinal vessels and five cells arranged in a circular pattern which may be a vascular bundle. In tangential section (Fig. 8.8d-f) tall longitudinal rays with 28 cells (Fig. 8.8c) can be seen, many of these exhibit homogenisation and slight splitting of the previously layered cell walls ('checking'; Jones, 1993) (Fig. 8.8e). Specimen G. 50242 is a fragment of partially charred fossil wood (Fig. 8.7i-k; T. Jones, pers. comm., 1994). This specimen exhibits outer tissues of secondary wood which have been burnt and are preserved as charcoal, and a slow transition from heat-affected 'charred' tissue to homogeneous, structureless lignites (Fig. 8.7i). The coalified tissues represent the inner core of the original wood fragment which were not burnt or affected by the heat of the fire and have subsequently undergone normal coalification. Similar fusain transition fossils have been noted elsewhere by Jones *et al* (1993).

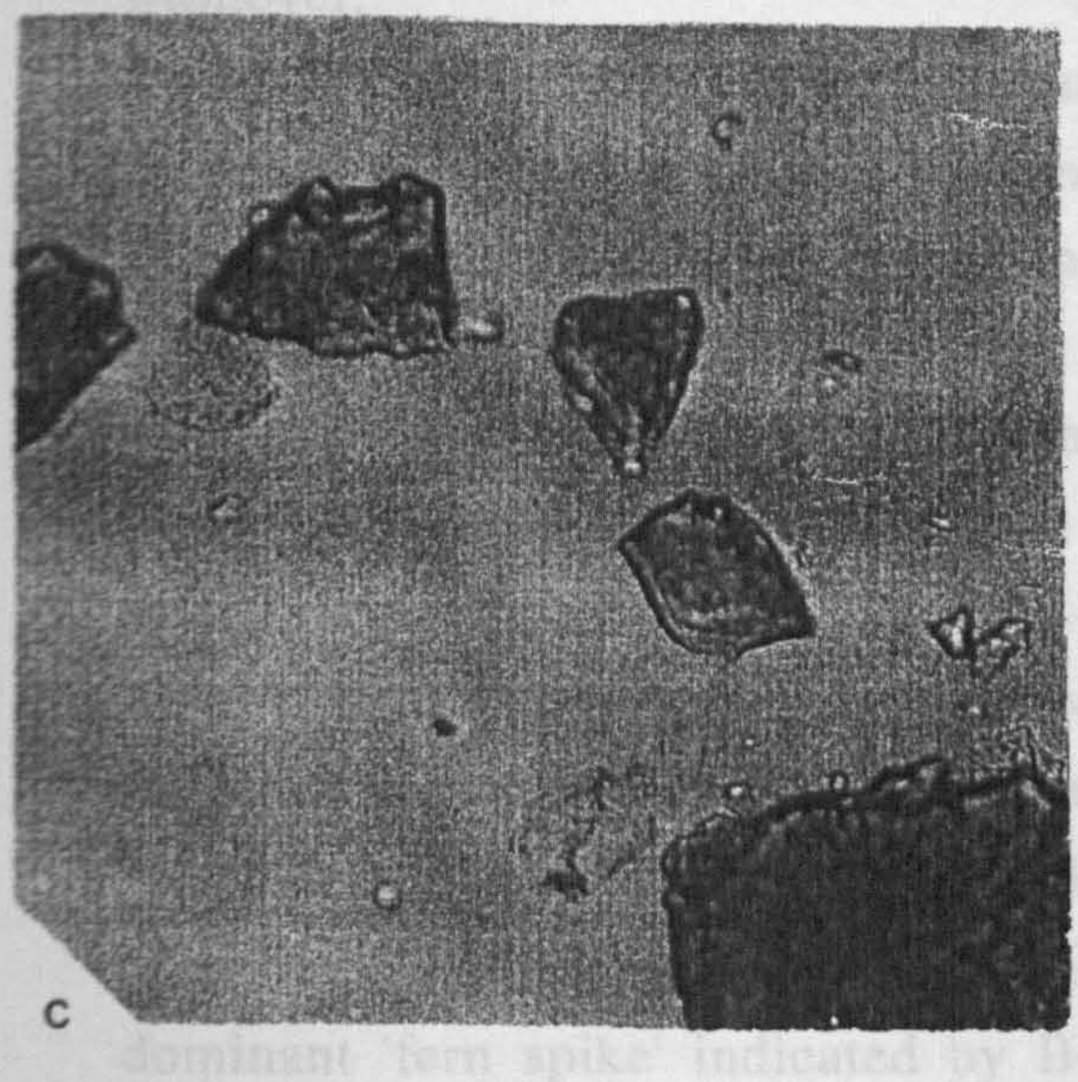
Close-up inspection of the charcoalled tissues reveals broken tracheid pits and homogenised cell-walls within specimen G. 50242 (Fig. 8.7k). Specimen G. 50926/7 also exhibits homogenised cell-walls (Fig. 8.8c,e) which indicates that these charcoal fragments were formed at a temperature above 300°C (Cope, 1980). Neither specimen shows much evidence of heat distortion, except for the large amount of cracking seen in the cycadophyte wood (Fig. 8.8f). This cracking might be related to shrinkage and expulsion of volatiles during charring of the outer wood (cf. high temperature cracking and 'checking' features in Jones, 1993) or may be a purely post-burial taphonomic feature. The palaeoecological significance of charcoal particles within the paleosols is considered within the next section.

(c) *Preservation of hard parts - spores and pollen.* Preservation of hard parts in an essentially unaltered state is seen in palynomorphs and cuticles (Spicer, 1991). This is because the walls of pollen grains and spores are composed of an extremely decay-resistant material called 'sporopollenin' (Spicer, 1991). 'Cutin', the material which is the major component of cuticles (Fig. 8.9d), is also remarkably resistant to degradation in an anaerobic environment. The preservation of the palynomorphs within the Grey Clay is extremely good (Figs. 8.6 and 8.9a). Some of the palynomorph cuticles exhibit slight degradation (Fig. 8.6j) or breakages (Fig. 8.6i,m). The excellent preservation of the organic cuticle (Fig. 8.9d) and tree resin (as resinite; Fig. 8.9c) indicates that the Grey Clay was deposited under anoxic or slightly acidic conditions within the pond (Metcalf *et al.*, 1992).

Within the upper Green Clay horizons the preservation of the spores and pollen is much poorer, indicating that the Green Clay represents deposition within relatively oxidising conditions that resulted in rapid degradation of organic material. This confirms that the impoverishment that is seen in the Green Clay palynomorph assemblages may be largely due to taphonomic rather than ecological reasons.







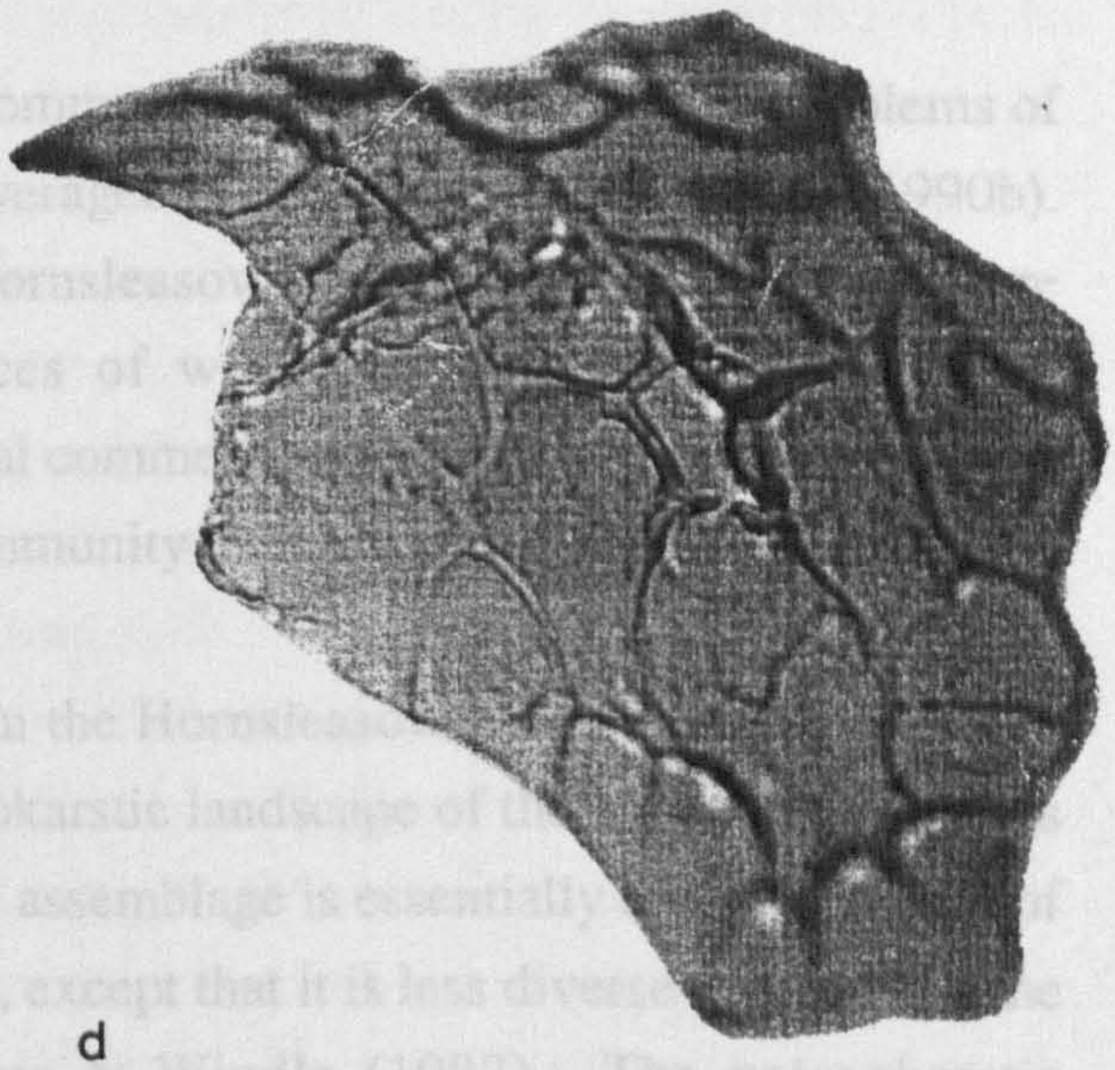


Figure 8.9. Preservation of plant microfossils, after Cole (1989). (a) Palynomorph-rich assemblage (i) with very little dispersed kerogen (ii) and some cuticle debris (iii) from the Grey Clay (1007) in grid square 3C (x100); (b) unstructured inertinite from 1007, grid 4D (x1000); (c) resinite from 1007, grid 3C (x500); (d) cutinite from 1007, grid 3C (x500).

(d) *Casts and moulds - ferric oxides and limonitised rootlets.* The ferruginous rhizocretions have been described above. They essentially form ferric oxide internal moulds where the original organic rootlet has rotted away. In most instances they are simply hollow or contain remnants of carbonised organic material, although some show internal radiating spurs (Fig. 8.7f), which probably correspond to original organic structures. The other forms of limonitised and ferric oxide replaced woody material are poorly preserved as casts of the original organic matter (Fig. 8.7c-e), these are either oxidised pyritised forms or associated with lignites.

8.4. Non-vertebrate palaeoecology and palaeoclimatic indicators for the British Bathonian

The features of the indigenous invertebrate fauna within the Hornsleasow clays are those of a low-diversity and low-abundance epifaunal palaeocommunity which developed within a low-nutrient, freshwater pond, with occasional brackish influences. Some mixing of the fauna also occurred from contemporary marine assemblages, by storm washover.

The reconstruction of ancient plant communities are fraught with the problems of dealing with ecologically mixed and time-averaged fossil accumulations (Scott, 1990b). This is especially true in the case of the Hornsleasow clays where *in situ* rootlets are associated with large allochthonous pieces of wood and a wind-or water-borne palynofloral fraction. However, a few general comments can be made about the nature of the English Midland coastal plain plant community as represented by the Hornsleasow assemblage.

Firstly, the floral remains seen within the Hornsleasow paleosols suggest that in the lower Bathonian the lower coastal palaeokarstic landscape of the northern Cotswolds was quite lushly vegetated. The palynofloral assemblage is essentially the same as that of other Bathonian localities in the British Isles, except that it is less diverse and without the dominant 'fern spike' indicated by Boulter & Windle (1993). The palaeokarstic floodplain areas would appear have been dominated by pteridophyte and lycopodaceous flora, similar to those of the Scalby Formation (Ravenscar Group: Bathonian) alluvial deposits in Yorkshire (Cope, 1993). Dryer areas of the floodplain would have supported forest-type vegetation, such as conifer, ginkgophytes and cycadophytes, but it is likely that some species of these groups would have been tolerant of swampy conditions (cf. *Taxodium distichum* or swamp cypress; Cope, 1993). This is supported by the presence of the large pieces of coniferous wood and the cycad charcoal fragment in the clays, as well as the palynofloral evidence.

The depositional environment of the Grey Clay paleosol was within fairly alkaline marshy-type conditions (section 5.12). The preservation of much of the invertebrate and plant remains, as ferruginous replacement and ferroan calcitic infills, points to a fairly low nutrient fresh water body. The occasional pyritised fossil and the preservation of

lignites in the Grey Clay, suggests that conditions may tended towards anoxia in the basal parts of the pond and within the underlying soils. Conditions may have become locally acidic by the build-up of humic acids and other organic decay products, which would have provided excellent preservation for plant fossils. This is a similar situation to soils which form in modern-day karstic dolines (e.g. in the Yorkshire Dales); these can become extremely acidic although surrounded by limestone and can in fact be a contributing factor in enlarging solutional hollows in the limestone (J. Cole, pers. comm). The physical damage and corrasion of shells, may have occurred within the pond or after burial. Small vertebrate remains also exhibit pitting and pock-marking by the action of humic and soil acids (section 10.4.2).

The presence of the *Equisetites* rootlets within the Green Clay, and the mottled and suncracked 'A' horizon at the top, are strongly suggestive that the boggy-freshwater environment within the doline underwent a period of drying out. In the Middle Jurassic, lycopodaceous plants were commonly the pioneering floras within marshy ground (for example, the *in situ* rootlet horizons seen in newly deposited floodplain muds and silts in the Ravenscar Group of Yorkshire; Cope, 1993). Although the diversity of the plant assemblage is much less within Green Clay, this may be largely because of the poorer preservational conditions within the soil. It is fairly clear that the change in environment from pond to soil, seen in Green Clay times (section 5.12), was a fairly localised event and caused few, if any, adverse effects to the plant community of the floodplain.

The presence of charcoal within the clays is interesting, as many other British Middle Jurassic coastal floodplain assemblages such as those of the Ravenscar Group (Yorkshire), the Brent Group (northern North Sea) and Brora Coal Group (Sutherland) also contain the charred remains of plants (Harris, 1958, Cope, 1993). This indicates that the coastal swamps in these regions were subjected to wildfires. Major groundfires have been recorded in modern-day and ancient peat-forming and swampy areas (Scott & Jones, 1993), and include those that occur within the Florida Everglades in periods of drought and warm weather (Cypert, 1972, Cohen, 1974). Wildfires are an integral part of modern ecosystems, and range from regular, sometimes seasonal and common occurrences in association with fire-adapted vegetation, to those which are rare, large and catastrophic to vegetation within wetter areas (Scott & Jones, 1993). Lightning is the principle cause of wildfire ignition (Komarek, 1972), but they can also be ignited by sparks created by rockfalls, volcanic activity, the spontaneous combustion of desiccated plant matter or burning marsh-gas (methane, CH_4). There appears to be no evidence for *in situ* sub-surface charring of rootlets or fossil peat (lignitic 'O' horizon) by ground fire within the Hornsleasow soils, which suggests that they were too wet.

If heavy storms may follow natural forest fires, there will be transportation and burial of charred matter. As charcoal is robust and floats in water, it can be easily transported over fairly large distances, before becoming waterlogged and incorporated in

the bedload sediment (Cope, 1993). The charred remains of the gymnosperm wood and the finer-grained inertinite fraction within the Hornsleasow deposits could have been transported and re-distributed by flooding and wind dispersal from drier, fire-prone regions of the floodplain such as forest and scrub. Hence, any palaeoecological conclusions about the nature of the Hornsleasow char are generalistic for the whole of the English Midlands coastal catchment area.

During the Bathonian the British Isles occupied a palaeolatitude position of between 30-35°N (Fig. 1.4). This is within a climatic area which is known as the 'Horse Latitudes' high pressure belt (Cope, 1993). The Middle Jurassic palaeocontinental configuration resulted in very little oceanic influence upon the climate of this region and the British Isle vegetation occupied an area of seasonal rainfall, generally known as the 'Seasonally Wet Belt' (section 1.3.3; Fig. 1.4). This area would have been subject to anticyclonic weather, ensuring a high atmospheric humidity, numerous thunderstorms and associated lightning during the monsoon season (Allen, 1975). During the dry season dead vegetation build up would ensure that there was much tinder for natural ignition. There seems to be much evidence that wildfire was a common component of British Bathonian terrestrial ecosystems (Harris, 1958, Cope, 1993).

Chapter Nine Vertebrate fauna

9.1. Vertebrates within the Chipping Norton Formation

The Chipping Norton Formation has long been associated with the recovery of dinosaur bones and the first recorded specimen of a dinosaur bone was discovered in 1677 at Cornwell, near Chipping Norton, Oxfordshire from the upper part of the Inferior Oolite succession (Plot, 1677), which was most probably equivilant to the Chipping Norton Formation. Although Plot correctly ascertained that the bone was from the thigh region, he thought that the bone may have belonged to an ox, horse or even "some elephant brought hither during the Government of the Romans in Britain" (Plot, 1677, p.131). It was later inaptly named *Scrotum humanum* (Brookes, 1763), but has recently been identified as the distal portion of a femur of the theropod dinosaur *Megalosaurus*, although the original specimen is now lost (Delair & Sarjeant, 1975).

Richardson (1911) refers to the fact that the limestone "has long been known for the occurrence in it or [within the] immediately superincumbent clay beds of the remains of the giant saurians of the genus *Cetiosaurus*" (Richardson, 1911, p. 204). *Cetiosaurus* bones were discovered at Chapel House (SP 329282), near Chipping Norton and communicated to the Geological Society of London in 1825, by Mr. J. Kingdon, who reported that the bones were from the 'upper portion' of the Bajocian 'Inferior Oolite' succession (Phillips, 1871, p. 245). The 'upper portion' of this sequence is, of course, the Chipping Norton Formation (Torrens, 1968): Beesley has also reported the presence of *Cetiosaurus* remains in the Inferior Oolite around Chipping Norton (Beesley, 1877, p. 185).

Other dinosaur remains found within the succession have been the theropod *Megalosaurus* (Richardson, 1911, 1929), and stegosaur plates refered to *Lexovisaurus vetustus* (Galton &Powell, 1983) have also been recovered from the Chipping Norton Formation at New Park Quarry, Gloucestershire. These were associated with an abundant vertebrate fauna including teleosaur crocodiles, turtles, dinosaurs and fish, which has been described by Richardson (1929), Gardiner (1935, 1937, 1939), Reynolds (1939), Galton & Powell (1983), Galton (1985). Other Chipping Norton Formation reptile sites have been fully documented by Benton (1988), Benton and Spencer (1995) and a list is given in Appendix A2.

9.2. The Hornsleasow vertebrate fauna (Table 9.1)

The Hornsleasow vertebrate assemblage consists of chondrichthyan and osteichthyan fishes, rare frogs and salamanders, a turtle, lizards and lepidosauromorphs, a

UNIT	BIOTIC ASSEMBLAGE	ECLGY	TAPHONOMIC ASSEMBLAGE	DERIVATION (in clays)	REFERENCE
Gy/Gn/Tr	Chondrichthyes	A(c)	To	*in situ* /reworked - CNF	T-69, M-92, B-94
Gy/Gn/Tr	Osteichthyes	A(c)	To/Sc/Ve	*in situ*	V-89, M-92, B-94
Gy/Gn	Amphibia	SA(fw)	Ve/Lb/J/Pl	*?in situ*	M-93, B-94
Gy/Gn	Reptilia				
Gy/Gn	Testudines	SA(c)	Ca/Ve	*in situ*	V-89, M-92, B-94
Gy/Gn	Lepidosauromorpha	T	Ja/Lb/Ve	local	V-89, M-92, B-94
Gy/Gn	Choristodera	SA(c)	Ja/Ve	*in situ*	M-92, B-94
Gy/Gn	Crocodylia: goniopholid	SA(c)	To/Da/Ve/Sk/Lb	*in situ*	V-89, M-92, B-94
Gy/Gn/Tr	teleosaur	A(m)	To	reworked - CNF	M-p, B-94
Gy/Gn	Pterosauria	T-airborne	To/Lb/Ve	local	V-89, M-92, B-94
Gy/Gn	Saurischia	T	To/Pc-undiff.*	local/derived/*in situ**	B-77, P-78, R-11*
Gy/Gn	Ornithischia	T	To	local/derived	V-89, M-92, B-94
Gy/Gn	Therapsida	T	To/Ve/Lb	local	V-89, M-92, B-94
Gy/Gn	Mammalia	T	To	local	V-89, M-92, B-94

Table 9.1. Vertebrate assemblage recovered from the Hornsleasow clays, showing palaeoecology, taphonomy and possible derivation. Abbreviations as for Tables 8.1. and 8.2; except 'To' teeth, 'Sc' scales, 'Ve' vertebrae, 'Lb' limb bones, 'Ja' jaws, 'Gi' girdle bones, 'Da' dermal armour, 'Ca' carapace, 'Sk' skull, 'Pc-undiff' undifferentiated post-cranial remains.

choristodere, various crocodilians, pterosaurs, and dinosaurs, a tritylodont, and rare teeth of several mammals. Particularly early records include the oldest frog of modern aspect and the earliest salamanders. The Hornsleasow vertebrate fauna has been described in full by Benton *et al.* (1995) and therefore the components of the assemblage shall only be briefly listed here, and their palaeoecological significance discussed in the next section. The components of the fauna can be placed into two major categories: (1) those elements obviously derived from the underlying limestones and surrounding rocks; and (2) those elements which form a component of the palaeokarstic environment. (This latter category is further subdivided by means of taphonomic and palaeoecological methods in the suceeding chapter). The vertebrates are listed and described systematically, based upon the classification used in Benton *et al.* (1995). This paper was the product of collaborative work between several authors on the identification and descriptions of the Hornsleasow taxa and their input is indicated at the beginning of the appropriate sections. All specimen 'G' numbers are prefixed by the abbreviation GLCRM (omitted here for brevity), which stands for Gloucester City Museum geological collections.

9.2.1. Derived vertebrate fauna.

These are commonly associated with much oolitic limestone and derived invertebrate matter, often broken and abraded and may have ooliths adhering to the surface. This element of the fauna was described by M.J. Benton and the author, and includes:

(a) *Chondrichthyes.* Specimens G. 50810 and G.50811 represent fragments from the very large lateral grinding teeth of the marine acrodont shark *Asteracanthus.* The original teeth would have been about 16 millimetres long and would have come from a beast of estimated 2-4 metres in overall length. They are clearly derived.

(b) *Reptilia.* Specimen G. 52729 is a worn and broken teleosaur crocodile tooth. This again is essentially a marine creature, and the tooth is clearly derived as limestone and ooliths are seen adhering to the broken surface.

9.2.2. The indigenous element

These are essentially an *in situ* aquatic and semi-aquatic fauna, with other components becoming incorporated within the pond by attrition. As with many similar microvertebrate assemblages much of the material is quite broken (Chapter 10) and identification is therefore limited in many cases. The assessments of relative abundance of each taxon by the various authors involved in the project are approximations based upon identifiable material and should not be considered to be a strict indication of palaeoecology. These considerations are outlined in more detail in the next chapter. The animals which comprise the indigenous fauna are listed systematically and described below, based upon the accounts given in Benton *et al* (1995).

9.3. Classification and general descriptions

9.3.1. Chrondrichthyes

(i) *Polyacrodontidae.* Five specimens of tiny narrow multiple-cusped shark teeth are present identified by M.J. Benton as *Polyacrodus* sp. Specimen G.66749 (Fig. 9.1a) is a tiny incomplete tooth, measuring 2.5mm in length. There is a little-worn triangular main cusp and three smaller lateral cusplets which diminish in size towards the extremity of the tooth. Specimen G.50031 (Fig. 9.1b), measuring 3mm in length, and lacks the ends. Both would have been 5-6mm long when complete. Unfortunately, both teeth largely lack the root protion which would probably have been broad and cancellous.

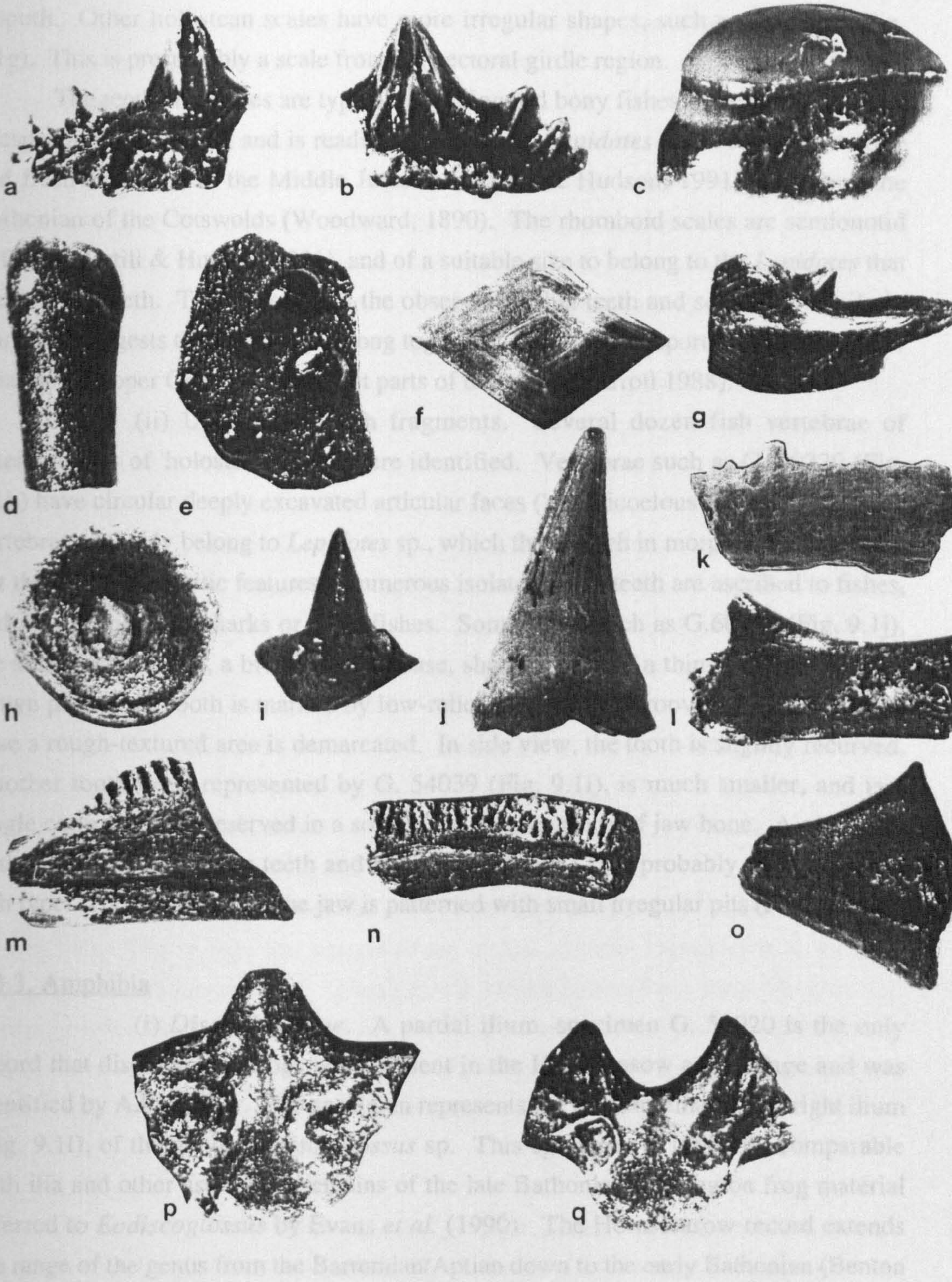
The teeth are typical grasping-type teeth of hybodont sharks, although they are very small. Illustrated specimens of *Polyacrodus brevicostatus* Patterson, 1966 (Cappetta 1987, fig. 40O-P) are similarly tiny, but bear a different sculpture. *Polyacrodus* has traditionally been regarded as a Triassic species, but Jurassic and Cretaceous material has been referred to this taxon (Cappetta 1987, p. 37). The present specimens are also similar to teeth of *Hybodus minor* Agassiz, 1837 figured by Duffin (1993, figs 11, 12) and Duffin and Delsate (1993, pl. 1, fig. 3) from the Late Triassic of England and Luxembourg: longer lateral teeth show the same proportions, the same shape of the central cusp, up to three lateral cusplets, a comparable linear and slightly sinuous ridge system on the cusps and similar size. If the present teeth are *Polyacrodus*, the record is unremarkable, but if they belong to *Hybodus minor*, this provides a range extension of a previously Triassic species.

9.3.2. Osteichthyes

(i) *Semionotidae.* Isolated crushing teeth, and abundant rhomboid scales are assigned to the semionotid genus *Lepidotes* sp.. Larger scales, such as those figured here (Fig. 9.1f), are rare, but hundreds of small thin rhomboid scales, measuring 1-2mm in length, also occur. Crushing teeth of *Lepidotes* type, such as G.50420 (Fig. 9.1c) are isolated from the original array of 10-20 which lay on broad bones in the palate. Another *Lepidotes* tooth, G.54840 (Fig. 9.1d) has a narrow cylindrical root set off from the spatulate blunt crown by a narrow groove. The specimen is 1.7mm long. One tiny palate specimen, G.50032 (Fig. 9.1e), shows seven irregularly arranged, crushing teeth deeply set in bone. The toothplate is small and quite fragile, although it seems to be relatively good condition and is possibly derived from a juvenile.

Holostean flank scales, such as G.52237 and 91202 (Fig. 9.1f) are robust, thick rhomboid elements. They display concentric growth lines and are dark brown in

Figure 9.1. Hornsleasow indigenous vertebrate fauna: Fishes and amphibians. Sharks: (a) G. 66749, partial *Polyacrodus* tooth, labial view (crown width: 2.5mm); (b) G. 50031, partial *Polyacrodus* tooth, lingual view (crown width: 3mm). Semionotids: (c) G. 50420, *Lepidotes* tooth, lateral view (crown width: 5.8mm); (d) G. 54840, *Lepidotes* tooth with root, lateral view (height: 1.7mm); (e) G. 50032, *?Lepidotes* tooth plate, occlusal view (1.7mm across). 'Holosteans': (f) G. 91202, flank scale, external view (7mm across); (g) G. 50060, pectoral scale, external view (9mm across); (h) G. 50720, amphicoelous vertebra, anterior view (3mm central diameter). Unidentified fish material: (i) G. 54039, tooth and partial jaw bone, lateral view (length: 1mm across jaw plate); (j) G. 60020, tooth, ?lingual view (height: 2.5mm); (k) G. 60813, partial jaw, labial view (length: 1.7mm across jaw plate). Amphibians: (l) G. 51020, *Eodiscoglossus* right ilium fragment, lateral view (length: 4.8mm); (m) G. 57747, *Marmorerpeton kermacki* dentary fragment, lingual view (length: 7mm); (n) G. 76910, *Marmorerpeton kermacki* dentary fragment, lingual view (length: 9mm); (o) G. 50709, *Marmorerpeton kermacki* humerus distal fragment, lateral view (length: 5.2mm); (p) G. 513526, 'Kirtlington Salamander B' atlas vertebra, anterior view (1.4mm across); (q) G. 58246, albanerpetonid atlas vertebra, dorsal view (1.3mm across).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colour. The outer face of the scale is flat, while the internal face is rounded and smooth. Other holostean scales have more irregular shapes, such as G.50060 (Fig. 9.1g). This is presumably a scale from the pectoral girdle region.

The teeth and scales are typical of semionotid bony fishes. The crushing tooth is especially diagnostic, and is readily assignable to *Lepidotes* sp., a common marine and freshwater fish in the Middle Jurassic (Martill & Hudson, 1991), including the Bathonian of the Cotswolds (Woodward, 1890). The rhomboid scales are semionotid in form (Martill & Hudson, 1991), and of a suitable size to belong to the *Lepidotes* that yielded the teeth. These facts, and the observation that teeth and scales are similarly abundant suggests that they may belong together. *Lepidotes* is reported from the Upper Triassic to Upper Cretaceous of most parts of the world (Carroll 1988).

(ii) Unidentified fish fragments. Several dozen fish vertebrae of osteichthyans of 'holostean' grade were identified. Vertebrae such as G. 50720 (Fig. 9.1h) have circular deeply excavated articular faces ('amphicoelous' condition). These vertebrae probably belong to *Lepidotes* sp., which they match in morphology and style, but they lack diagnostic features. Numerous isolated small teeth are ascribed to fishes, and may come from sharks or bony fishes. Some teeth, such as G.60020 (Fig. 9.1j), are small and conical, a broad, hollow base, showing part of a thin-walled root. The crown part of the tooth is marked by low-relief longitudinal grooves, and towards the base a rough-textured area is demarcated. In side view, the tooth is slightly recurved. Another tooth type, represented by G. 54039 (Fig. 9.1i), is much smaller, and is a single conical tooth preserved in a socket in a slender piece of jaw bone. A small jaw with 2-3 broken, peg-like teeth and 8 sockets (G. 60813) is probably from a similar fish type. The labial side of the jaw is patterned with small irregular pits (Fig. 9.1k).

9.3.3. Amphibia

(i) *Discoglossidae*. A partial ilium, specimen G. 51020 is the only record that discloglossid frogs were present in the Hornsleasow assemblage and was identified by A.R. Milner. The specimen represents the broken remains of a right ilium (Fig. 9.1l), of the genus *Eodiscoglossus* sp. This specimen is probably comparable with ilia and other associated remains of the late Bathonian Kirtlington frog material referred to *Eodiscoglossus* by Evans *et al.* (1990). The Hornsleasow record extends the range of the genus from the Barremian/Aptian down to the early Bathonian (Benton *et al.*, 1995). The Kirtlington material formed the basis of the new species *E. oxoniensis*. The Hornsleasow fragmentary ilium differs from those at Kirtlington in overall morphology (Benton *et al.*, 1995). Therefore, it is possible that the Hornsleasow frog is specifically distinct, but more complete material would be required to establish this.

(ii) *Caudata incertae sedis*. A partial large atlas vertebra (specimen G.58722) has been identified as that of the undifferentiated salamander *Marmorerpeton kermacki* by A.R. Milner and S.E. Evans. This species is previously recorded from the Upper Bathonian site at Kirtlington (Evans *et al.*, 1988). A range of other material including dentaries (specimens G. 57747 (Fig. 9.1m), 69905, 69906, 76901, 76629, 76904, 76907, 76909, 76910 (Fig. 9.1n), 76912, 76913, 76914, 76915, 76916), a maxilla (specimen G. 76920), a premaxilla (G. 76908), a caudal vertebra (G. 76624) and a humerus (G. 50709: Fig. 9.1o) can also be assigned to this genus (Benton *et al.*, 1995). However, it is not possible to attribute individual elements to one or other of the named species. *Marmorepeton* has been recorded from the late Bathonian of Dorset and Oxfordshire (Evans *et al.*, 1988, Evans, 1993) and from the Oxfordian/Kimmeridgian lignite beds of Guimarota, Portugal (Evans, pers. obs., *in* Benton *et al.*, 1995). The Hornsleasow material therefore represents the earliest record of *Marmorerpeton* and hence, also of salamanders in general.

(iii) *Caudata incertae sedis*. Specimen G. 50846 is a small atlas vertebra centrum of a type that is distinct from that of *Marmorerpeton* (Benton *et al.*, 1995) and is similar to the "Kirtlington Salamander A" (currently under study by S.E. Evans and A.R. Milner, but as yet undescribed) which is represented in that fauna by hundreds of specimens.

(iv) *Caudata incertae sedis*.Specimen G. 53526 is a tiny atlas vertebra centrum of a type distinct from those described above (Fig. 9.1p). It is similar in type to the derived, but as yet undescribed "Kirtlington Salamander B" remains, which are again under study by S.E. Evans and A.R. Milner.

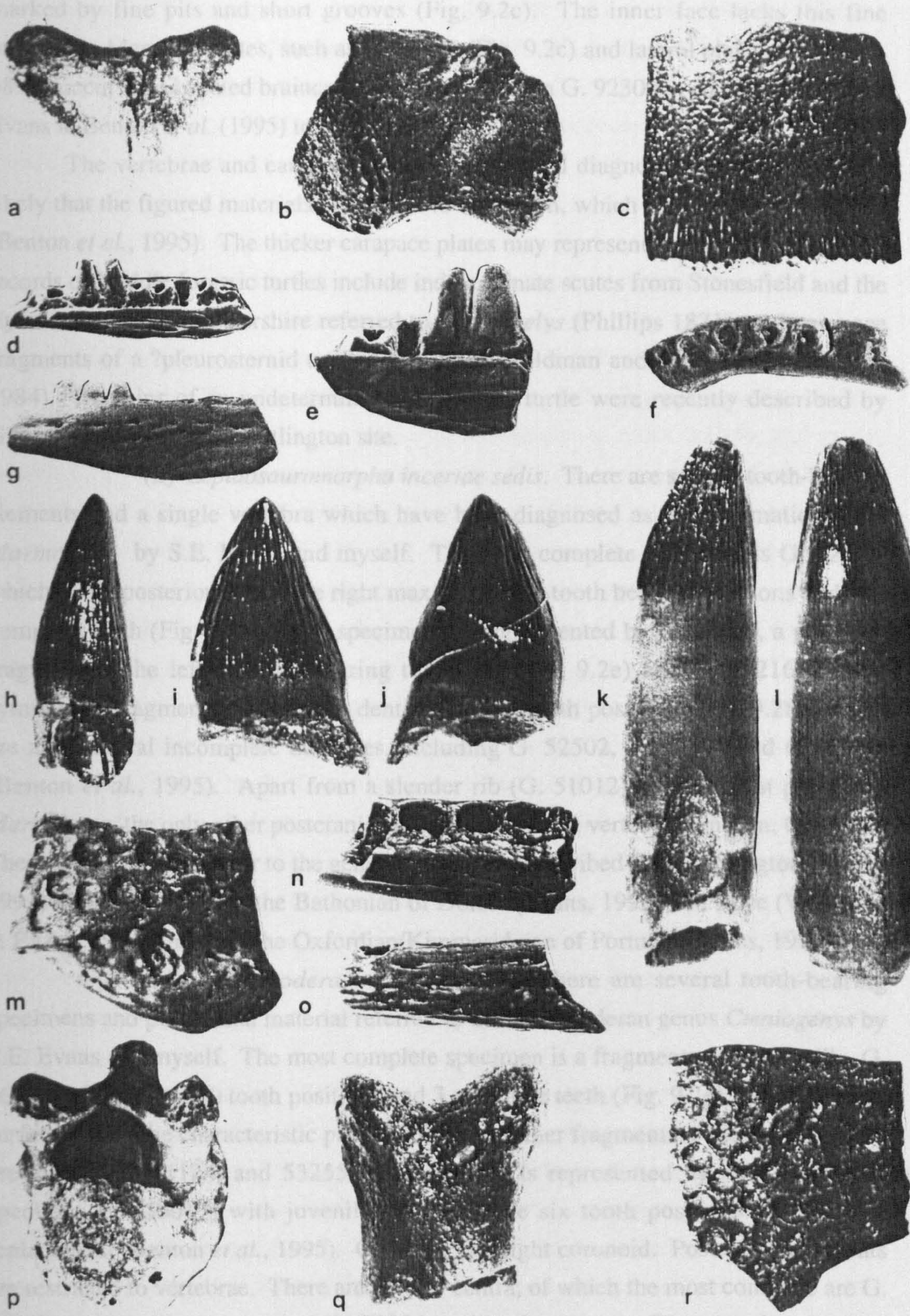
(v) *Amphibia incertae sedis*. Specimen G.58246 is a small and unsual atlas element (Fig. 9.1q) The unique shape of this element identifies it as an atlas vertebra of an albanerpetontid. These small amphibians have a long fossil record extending from the Bajocian to the Miocene, and a wide geographic range including North America, Europe and Central Asia (A.R. Milner and S.E. Evans, *in* Benton *et al*, 1995). They are of uncertain systematic position.

9.3.4. Reptilia

(i) *Pleurosternidae*. Several partial vertebrae, and hundreds of carapace fragments identified by S.E. Evans and M.J. Benton, most of them lack margins and suture lines, and many are heavily abraded. Vertebrae include fragments of the neural arch, G. 76608, 76612 (Fig. 9.2a), a partial caudal vertebra, G. 92501 (Fig. 9.2b) and a more complete turtle caudal vertebra G. 66759 measuring only 4mm long.

Most carapace and plastron fragments are incomplete, although some show marginal features. Plates are generally 2-3mm thick, although some abraded larger 5-10mm thick turtle carapace fragments also occur. The outer surface is smooth and

Figure 9.2. Hornsleasow indigenous vertebrate fauna: Turtles, crocodiles and small reptiles. Pleurosternid turtles: (a) G. 76612, neural arch fragment, ventral view (4.1mm across); (b) G. 92501, partial caudal vertebra, lateral view (length: 4.0mm); (c) G. 55920, carapace fragment, external view (11.7mm across). Lepidosauromorphs: (d) G. 52640, *Marmoretta* posterior end of right maxilla, lingual view (length: 4.4mm); (e) G. 66854, *Marmoretta* posterior fragment of left maxilla, lingual view (length: 2.8mm); (f) G. 66855, *Marmoretta* symphysial fragment of right dentary, occlusal view (length: 2.3mm). Choristoderes: (g) G. 66855, *Cteniogenys* left maxilla fragment, labial view (length: 4.4mm). Goniopholid crocodiles: (h) G. 51201, tooth type-1 with partial root, lingual view (height: 4.5mm including root); (i-j) G. 51415, tooth type-2 with partial root, labial (i) and lingual (j) views (height: 2mm including root); (k-l) G. 75216, tooth type-1 with root and basal replacement pit (k), lingual (k) and labial (l) views (height: 13.5mm including root); G. 75101, sculptured squamosal, external view (length: 9.5mm); (n-o) G. 51525, jaw fragment, occlusal (n) and labial (o) views (length: 12mm); (p) G. 66759, vertebra, posterior view (4.1mm across); (q) G. 51034, distal fragment of phalanx, anterior view (length: 3mm); (r) G. 75401, sculptured scute fragment, external view (10mm across).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marked by fine pits and short grooves (Fig. 9.2c). The inner face lacks this fine ornament. Marginal plates, such as G. 55920 (Fig. 9.2c) and lateral plates, such as G. 58937 occur. An isolated braincase element, specimen G. 92302 is considered by S.E. Evans in Benton *et al.* (1995) to be from a turtle.

The vertebrae and carapace plates offer limited diagnostic information. It is likely that the figured materials all pertain to one taxon, which may be a pleurosternid (Benton *et al.*, 1995). The thicker carapace plates may represent another taxon. Earlier records of Middle Jurassic turtles include indeterminate scutes from Stonesfield and the Eyford area of Gloucestershire referred to *Protochelys* (Phillips 1871), and carapace fragments of a ?pleurosternid turtle from Skye (Waldman and Savage 1972; Savage 1984). Remains of an undetermined cryptodiran turtle were recently described by Gilham (1994) from the Kirtlington site.

(ii) *Lepidosauromorpha incertae sedis.* There are several tooth-bearing elements and a single vertebra which have been diagnosed as the enigmatic reptile *Marmoretta* by S.E. Evans and myself. The most complete specimen is G. 52640, which is the posterior end of the right maxilla, with 9 tooth bearing positions and two complete teeth (Fig. 9.2d). This specimen is supplemented by G. 66854, a posterior fragment of the left maxilla, bearing two teeth (Fig. 9.2e) and G. 58216, a right symphysial fragment of a juvenile dentary, with 7 tooth positions (Fig. 9.2f). There are also several incomplete dentaries, including G. 52502, G. 52134 and G. 76628 (Benton *et al.*, 1995). Apart from a slender rib (G. 51012) which might pertain to *Marmoretta*, the only other postcranial element is a dorsal vertebral centrum, G. 76627. These elements are similar to the genus *Marmoretta* described from Kirtlington (Evans, 1991) and recorded from the Bathonian of Dorset (Evans, 1993) and Skye (Waldman & Evans, 1994) and from the Oxfordian/Kimmeridgian of Portugal (Evans, 1991).

(iii) *Choristodera incertae sedis.* There are several tooth-bearing specimens and postcranial material referred to the choristoderan genus *Cteniogenys* by S.E. Evans and myself. The most complete specimen is a fragmentary left maxilla, G. 66855, which bears 10 tooth positions and 3 complete teeth (Fig. 9.2g), and the lateral surface shows the characteristic pitted sculpture. Other fragmentary pieces of maxilla are G. 51035, 51124, and 53255. The dentary is represented by only one small specimen (G. 50017) with juvenile dentition, the six tooth positions show active replacement (Benton *et al.*, 1995). G. 92301 is a right coronoid. Postcranial elements are restricted to vertebrae. There are several centra, of which the most complete are G. 76808, a cervical centrum and G. 92304, a dorsal centrum. The neural arches are never well preserved, but include G. 76802 and G. 76803, both of which show the base of the neural arch and parts of the posterior zygapophyses.

The jaw and vertebral elements described by S.E. Evans in Benton *et al.* (1995) closely match similar material from Kirtlington (Evans, 1990a, b) referred to the genus

Cteniogenys. *Cteniogenys* was first described from the Upper Jurassic (Kimmeridgian/Tithonian) Morrison Formation of North America (Gilmore, 1928), as a possible lizard. However, a more comprehensive study by S.E. Evans of the material from Kirtlington has shown it to be a small choristodere (Evans, 1990a, b). *Cteniogenys* has been recorded from the late Jurassic of Portugal (Guimarota) and from other Bathonian localities in Britain (Evans, 1993). However, these specimens are the earliest record of this genus.

(iv) *Goniopholidae*. Crocodiles are the most abundant elements of the Hornsleasow fauna, being represented mainly by teeth, of which several hundred have been recovered. There are rarer skull elements, identifiable postcranial elements, and scutes. These were identified by M.J. Benton and myself. Typical posterior teeth of 'goniopholid' type (Fig. 9.2h) are circular to subcircular in cross section, and gently recurved. The labial surface of the tooth is set off from the lingual surface by very slight keels which which may show pseudoserrations (Benton *et al.*, 1995) for example specimens G. 90495, 90497, 52319, and 90519. The enamel covering the crown of the tooth is generally smooth, except on the concave face where slight vertical striations may develop. These striations develop into slight ridges and grooves towards the base of the tooth. There are also many hundreds of tiny teeth of the same shape, but only 1-2mm long. These may be teeth of juveniles, or barely erupted teeth from older animals.

Another tooth morphology is rather rarer, but may also be 'goniopholidid' and may represent anterior teeth of the same animal that produced the teeth just described. The second tooth morphology consists of shorter teeth, with crowns 2-5mm long (Fig. 9.2i,j). A complete specimen, G. 75216 (Fig. 9.2k,l), has a long root which shows a resorption pit low down on the internal side, where it presumably lay close to the root of an adjoining tooth. The crown is broad and compressed from side to side. Other specimens of crowns, such as G. 51415 (Fig. 9.2i,j), show the striation pattern on the lingual face very clearly and, in some cases, a clear constriction between the crown and the root.

Certain crocodilian skull elements may be recognised. Part of the left squamosal, G. 75101 (Fig. 9.2m) shows the characteristic crocodilian sculpture of circular pits on the external surface, and the curved margin of the posterior angle of the skull table. A jaw fragment with deep alveoli, and some remnants of the roots of teeth that are circular in cross-section, G. 51525 (Fig. 9.2n) is probably also crocodilian. The lingual and ventral faces (Fig. 9.2o) bear numerous blood vessel and nerve openings.

Identifiable crocodilian vertebrae are surprisingly rare, but imclude partial centrums, G. 75501, G. 76801 and G.76809 and more complete vertebrae, G. 66759 (Fig. 9.2p). One or two phalanges may be crocodilian: a partial distal end, G. 51034 (Fig. 9.2q) shows the articular portion which fits into a concave socket on the next

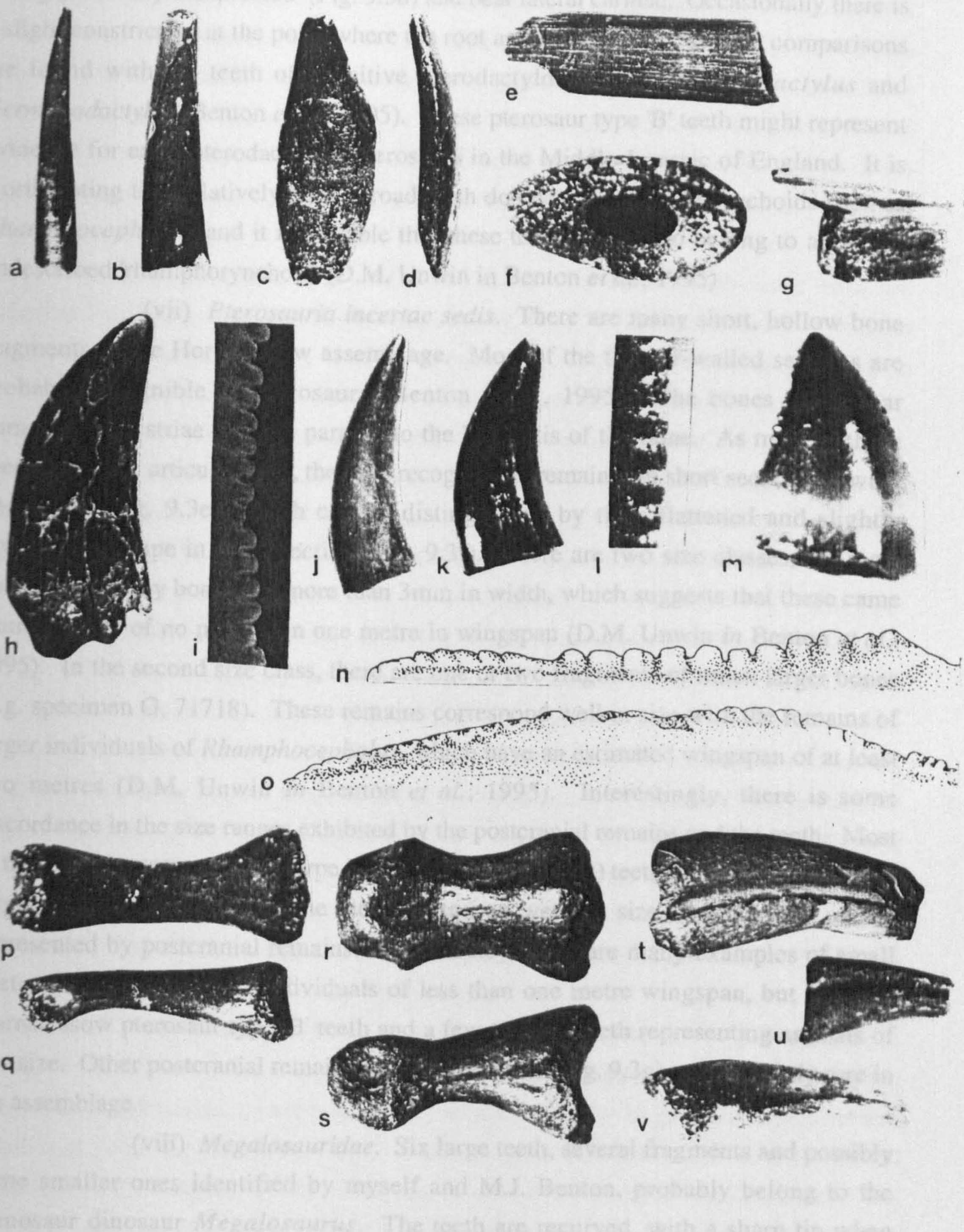
most distal phalanx. After teeth, the commonest crocodilian remains are scute fragments. These are all 1-2mm thick, and they show the characteristic crocodilian ornament of numerous closely-packed circular pits on the external surface, and some show sharp angles or smooth marginal portions (Fig. 9.2r).

All the crocodilian material can be ascribed to *Goniopholis*, but with the understanding that this genus is somewhat of a waste-basket taxon that may include several distinct genera. The taxon *Goniopholis* is currently under review by J.B. Delair (pers. comm., 1994) and is considered by him to be strictly referable to only the Purbeck and Wealden occurrences (e.g. Ensom *et al.*, 1994) and perhaps the Portland material (J.B. Delair pers. comm., 1994). Delair suggests that "the Stonesfield Slate (*sic*) and earlier goniopholid remains may turn out to merit some distinctive new name" (pers. comm., 1994). Nonetheless, all the teeth, bones, and scutes, could belong to the same taxon, both in terms of morphology, and in terms of their range in sizes. The two tooth morphologies could be loosely identified as 'goniopholidid' (Fig. 9.2h) and 'atoposaurid' (Fig. 9.2i,j), but the latter teeth lack a key atoposaurid feature: the internal vertical ridges do not diverge from the marginal ridges (Evans and Milner, 1994), and instead seem to run roughly in parallel from the distal tip. The differing tooth morphologies most likely occur as part of individual variation within the jaw of the animal, as most crocodilians including the modern *Crocodylus* exhibit a degree of heterodonty. Goniopholidids range in age from the late Jurassic to latest Cretaceous, and perhaps from the mid Jurassic onwards, should these, and other (Evans and Milner, 1994) records from the Bathonian stage be confirmed.

(v) *Rhamphorhynchidae.* Pterosaurs are not uncommon in the Hornsleasow assemblage being represented by many teeth and numerous fragmented postcranial remains. Virtually all the teeth were assigned to one type by D.M. Unwin as Hornsleasow pterosaur type 'A' (Benton *et al.*, 1995), these are elongate, narrow, sharply pointed and gently recurved backwards (Fig. 9.3a). There is very little variation on this basic type and the shape, and 'V'-shaped enamel distribution (Fig. 9.3b) are typical of the members of the Rhamphorhynchidae. These are the only 'fang-toothed' pterosaur family known to be present in the Middle Jurassic, however, isolated teeth are not diagnostic. *Rhamphocephalus* is the only rhamphorhychine so far known fron the Middle Jurassic of Europe and has been recorded from a number of English Bathonian localities including Stonesfield (Huxley, 1859), Eyford (pers. obs.) and Kirtlington (Evans & Milner, 1994). The Hornsleasow pterosaur type 'A' is tentatively assigned to *Rhamphocephalus* by D.M. Unwin *in* Benton *et al.* (1995).

(vi) *Pterodactyloidea incertae sedis.* This tooth type (Hornsleasow pterosaur type 'B') is very rare with only six specimens (G. 51423, 52203, 57461, 66861, 66862, 92705) identified by D.M. Unwin so far. The teeth are short and broad with a slightly asymmetric triangular outline in labial aspect (Fig. 9.3c). They are quite

Figure 9.3. Hornsleasow indigenous vertebrate fauna: Pterosaurs and theropods. Pterosaurs: (a) G. 66841, Hornsleasow pterosaur-type 'A' (rhamphorynchid fang), labial view (crown height: 10.2mm); (b) G. 54311, Hornsleasow pterosaur-type 'A' with 'V'-shaped enamel distribution, lingual view (crown height: 10.5mm); (c-d) G. 57461, Hornsleasow pterosaur-type 'B' (pterodactyloid tooth), labial (c) and lateral (d) views (height: 6.9mm including root); (e-f) G. 55410, striated wing phalanx lacking articular ends, lateral (e) and cross-sectional (f) views (length: 5.7mm); (g) G. 76618, vertebra, lateral view (length: 6.5mm). Theropods: (h) GMHM1, *Megalosaurus* shed crown with tip-wear facet (crown height: 68.3mm); (i) G. 66724, *Megalosaurus* tooth posterior margin serrations, 4 per mm; (j) G. 51611, maniraptoran type-A shed crown (crown height: 5mm); (k) G. 50836, maniraptoran type-A shed crown (crown height: 4mm); (l) maniraptoran type-A posterior margin serrations, 13 per mm; (m-o) G. 51422, maniraptoran type-B shed crown (crown height: 5.5mm), with 4 serrations per mm upon the posterior margin (n) and 5 serrations per mm on the anterior margin (o); (p-q) G. 50013, phalanx, upper (p) and lateral (q) views (length: 10.5mm); (r-s) G. 66737, phalanx, upper (q) and lateral (s) views (length: 4.6mm); (t) G. 77400, claw, lateral view (length: 17.9mm); (u) G. 77301, claw, lateral view (length: 0.5mm); (v) G. 77305, claw, lateral view (length: 11mm).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strongly laterally compressed (Fig. 9.3d) and bear lateral carinae. Occasionally there is a slight constriction at the point where the root and crown meet. Closest comparisons are found with the teeth of primitive pterodactyloids such as *Pterodactylus* and *Germanodactylus* (Benton *et al.*, 1995). These pterosaur type 'B' teeth might represent evidence for early pterodactyloid pterosaurs in the Middle Jurassic of England. It is worth noting that relatively short, broad teeth do occur in rhamphorynchoids such as *Rhamphocephalus*, and it is possible that these teeth might also belong to an as yet undescribed 'rhamphorynchoid' (D.M. Unwin in Benton *et al.*, 1995)

(vii) *Pterosauria incertae sedis.* There are many short, hollow bone fragments in the Hornsleasow assemblage. Most of the thinner-walled sections are probably assignible to pterosaurs (Benton *et al.*, 1995). The bones often bear numerous fine striae running parallel to the long axis of the bone. As most of these specimens lack articular ends, the only recognisable remains are short sections of wing phalanges (Fig. 9.3e), which can be distinguished by their flattened and slightly asymmetric shape in cross section (Fig. 9.3f). There are two size classes, the most abundant are tiny bones, no more than 3mm in width, which suggests that these came from animals of no more than one metre in wingspan (D.M. Unwin *in* Benton *et al.*, 1995). In the second size class, there are one or two fragments of much larger bones (e.g. specimen G. 71718). These remains correspond well in size with the remains of larger individuals of *Rhamphocephalus*, which have an estimated wingspan of at least two metres (D.M. Unwin *in* Benton *et al.*, 1995). Interestingly, there is some discordance in the size ranges exhibited by the postcranial remains and the teeth. Most of the Hornsleasow pterosaur type 'A' (?*Rhamphocephalus*) teeth belong to a relatively large individuals of at least one metre wingspan, yet this size range is very poorly represented by postcranial remains. Converseley, there are many examples of small postcranial remains from individuals of less than one metre wingspan, but only six Hornsleasow pterosaur type 'B' teeth and a few type 'A' teeth representing animals of this size. Other postcranial remains such as vertebrae (Fig. 9.3g) are extremely rare in the assemblage.

(viii) *Megalosauridae.* Six large teeth, several fragments and possibly some smaller ones identified by myself and M.J. Benton, probably belong to the carnosaur dinosaur *Megalosaurus*. The teeth are recurved, with a sharp tip when unworn (Fig. 9.3h), and the lingual and labial faces are separated by sharp anterior and posterior ridges which bear serrations spaced at between 1.5-4 per mm depending on tooth size (e.g. specimens G. 66724, G. 66724, GMHM1) (Fig. 9.3i). These teeth are indistinguishable from those in the type specimen dentary fragment of *Megalosaurus bucklandi* Meyer, 1832 (OUM J12142) from Stonesfield. There are no clear-cut diagnostic characters of such teeth, and they should probably be designated simply as those of a 'carnivorous archosaur'. However they match those of *M. bucklandi*

precisely in size, shape, and distribution of serrations. The taxonomy of *Megalosaurus* is a nightmare, with species reported from Rhaetian (latest Triassic) to early Cretaceous (Molnar, 1990), most of which probably belong to different genera, and some even to different families. The type material of *M. bucklandi* at least must be identified acceptably, and the same may be the case for other material from the Bathonian of the Cotswolds, including the present specimens, which are from rocks dated as perhaps 5 Myr. older than those at Stonesfield.

(viii) *Maniraptora incertae sedis*. Between 30-40 small teeth are described in Benton *et al.* (1995) as Maniraptora Type A [?cf. ceratosaur]. The teeth have long, slender and slightly recurved crowns (Fig. 9.3j,k). The anterior margin, on the outside of the curve, is rounded in cross section, while the posterior margin comes to a narrow keel which bears small, regular serrations spaced 8-15 per mm (e.g. specimens G. 51036, G. 51820, G.60008, G. 92501) (Fig. 9.3l). For the majority of the teeth the anterior margin shows no serrations, except for a short series of tiny worn crenulations high on the curve, just below the tip. These teeth are identified tentatively as maniraptoran by comparison with specimens figured by Currie *et al.* (1990), and are similar to 'dromaeosaur-like' teeth from the British Bathonian illustrated by Evans & Milner (1994). It is unlikely that these teeth truly represent dromaeosaurid dinosaurs: the oldest relatively completely known dromaeosaurid taxon is *Deinonychus antirrhopus* Ostrom, 1969, from the Cloverly Formation of Wyoming and Montana, which is Aptian or Albian in age (mid Cretaceous), some 60 Myr. younger than the Hornsleasow deposits. However, maniraptoran dinosaurs are likely to occur in the mid Jurassic, since birds arose from within Maniraptora at least by latest Jurassic times (*Archaeopteryx*, Tithonian), and a possible maniraptoran jaw element, *Lisboasaurus*, has been reported (Milner and Evans, 1991) from the mid or late Jurassic (Callovian-Oxfordian) of the Guimarota locality in Portugal.

(ix) *Maniraptora incertae sedis*. There are between 10-15 small teeth of a second maniraptoran tooth type, identified as Maniraptora Type B, is distinguished from that just described by being generally smaller, labially-lingually compressed and more triangular in shape (Fig. 9.3m). Further differences are that both anterior and posterior edge margins of the teeth are invested with coarse serrations, and the serrations are spaced 3.5-5 per mm but becoming smaller towards the tooth tip and the base of the crown (Fig. 9.3n,o). The serrations are also much less regular in appearence, and have a hooked shape (e.g. specimens G. 51422, G. 60007, G. 56801, G. 50823, G. 60016, G. 60012). These teeth are identified tentatively as maniraptoran, and possibly of generally 'troodont' type, on the basis of comparisons with teeth illustrated by Currie (1987), Currie *et al.* (1990), Farlow *et al.* (1991), and following the interpretation by Evans and Milner (1994). These teeth probably do not come from theropods of the family Troodontidae, a family whose oldest confirmed

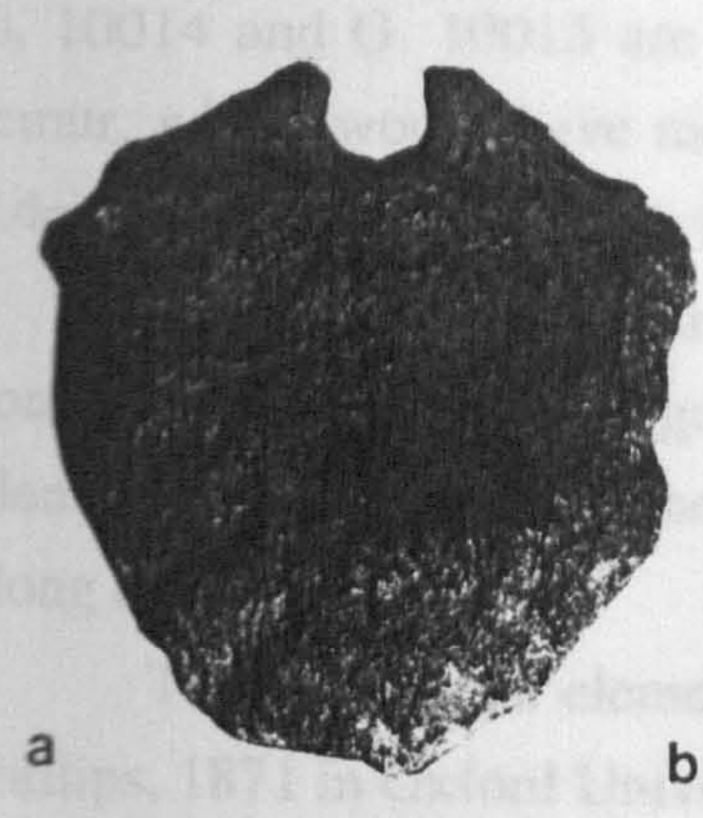
representative is *Saurornithoides mongoliensis* Osborn, 1924, from the Djadochta Formation of Mongolia, dated as Santonian or Campanian (late Cretaceous), 90 million years younger than the Hornsleasow deposits, but they may be from related animals.

(x) *Theropoda incertae sedis.* Twenty or thirty small phalanges and claws. Two phalanges, G.50013 (Fig. 9.3p,q) and G.66737 (Fig. 9.3r,s) are comparable, and may belong to a small theropod. A number of different types of claw phalanges occur. Larger ones, like G. 77400 (Fig. 9.3t), would have been about 20-25mm long when complete. The claw phalanx is gently curved, laterally compressed, and bears a deep channel running the full length of each side, presumably for anchoring the keratinous claw. A second type of claw phalanx, such as G. 77301 (Fig. 9.3u), is much smaller, measuring perhaps 7-8mm long when complete, and bearing the channel rather higher on each lateral face than in the first type. The third claw phalanx type, represented by G. 77305 (Fig. 9.3v), is larger, at 11mm long, and rather lower. The lateral channel is shorter, appearing only in the distal half of the phalanx, and occurring low on the claw. The phalanges were identified as those of a theropod by the presence of the deep ligament pits (Benton *et al.*, 1995). Their size indicates that they could have come from either the 'dromaeosaurid' or the 'troodontid'. The claw phalanges are harder to place. Perhaps claw types 1 and 2 are theropod, belonging to a small *Megalosaurus* and a maniraptoran respectively. Claw type 3 could be theropod or crocodilian (Benton *et al.*, 1995).

(xi) *Cetiosauridae.* Two teeth, 25 vertebrae (five complete) and 4-5 limb and girdle elements (2 fragmentary pelves, the femur), plus fragments of limb shafts, podia, ribs, 1 partial jaw fragment. The postcranial elements and jaw fragment most probably respresent the partial skeleton of one individual (Metcalf *et al.*, 1992). The elements are completely disarticulated, and largely disassociated from one another. The vertebrae comprise large fragments of centra from several regions of the axial skeleton (e.g. G. 10005 (Fig. 9.4a: ?caudal), G. 10006 - G. 10008, G. 10013, G. 10016 - G. 10019), measuring between 0.2-0.5 metre in diameter, these are associated with other fragmentary axial remains, including neural spines, an abraded neural complex (G. 10157) and other vertebral processes (e.g. G. 10137, G. 10155). More complete specimens also occur, and include several cervical neck vertebrae (e.g. G. 10003, G. 10009 - G. 10012, G. 10156) (Fig. 9.4b), some dorsal remains (e.g. G. 10000, G. 10124) (Fig. 9.4c) and caudal vertebrae (e.g. G. 10002) (Fig. 9.4d). Broken chevron bones (G. 10118, G. 10119, G. 10134) and several fragments of ribs were also recovred from the site.

There are three large pieces of hip girdle, including the right ilium with anterior peduncle and characteristically straight pubic process (G. 10001: Fig. 9.4e,f). The back of the pubic process is absent, so that it is not possible to see the shape and depth of the acetabulum, and much of the iliac blade is also missing (Fig. 9.4e,f). Specimens

Figure 9.4. Hornsleasow indigenous vertebrate fauna: cetiosaur sauropods: (a) G. 10005, caudal vertebra centrum, posterior view (central width: 214mm); (b) G. 10010, cervical vertebra, lateral view (antero-posterior width: 235mm); (c) G. 10000, mid-posterior dorsal vertebra, anterior view (central width: 200mm); (d) G. 10002, mid-anterior caudal vertebra, anterior view (central width: 290mm); (e-f) G. 10001, fragmentary right ilium blade with partial pubic process, lateral views (ilium blade length: 0.64m), with inset showing cartoon of position of fragment on the hip girdle; (g) G. 10020, sketch of restored femoral shaft (length: 1.5m); (h-i) G. 50057, tooth with partial root, lingual (h) and lateral (i) views (crown height: 14.3mm); (j) G. 75712, worn tooth, lingual view (crown height: 8.6mm).



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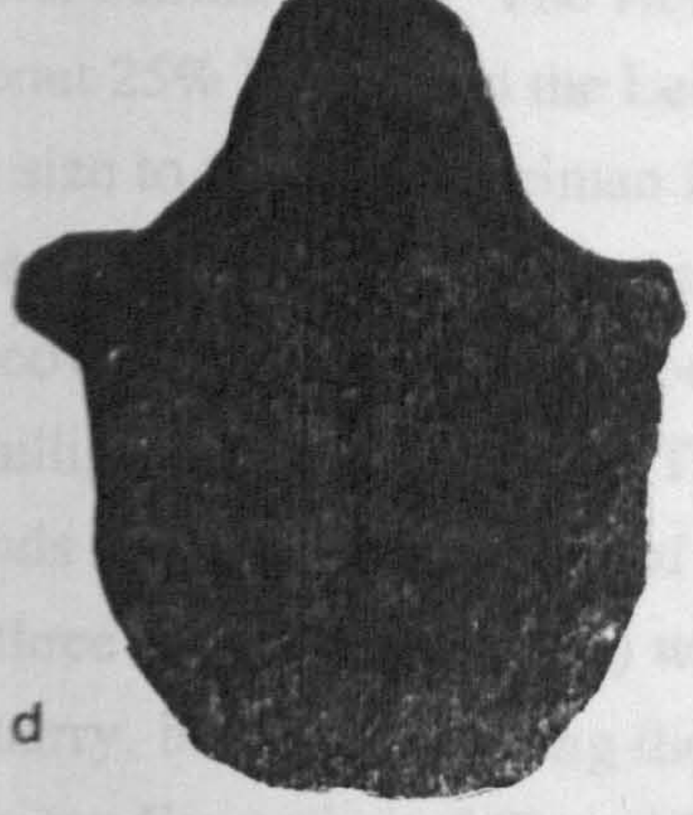

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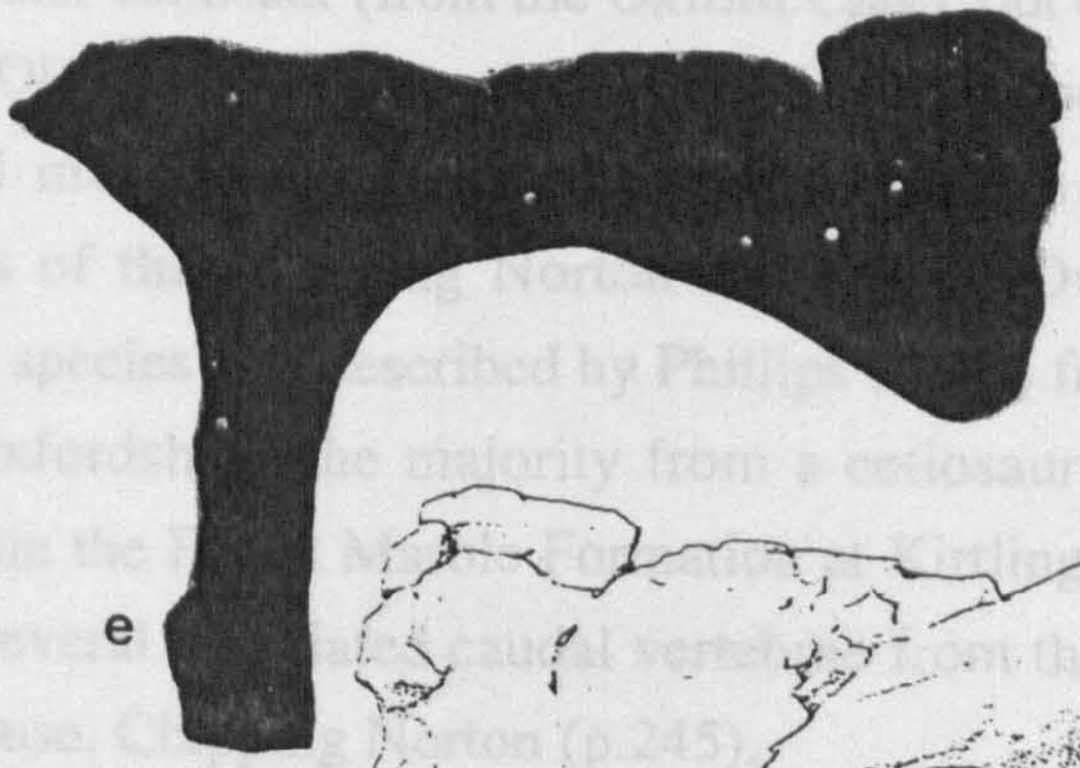

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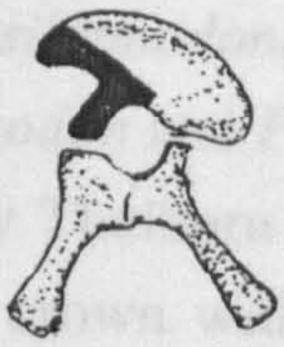

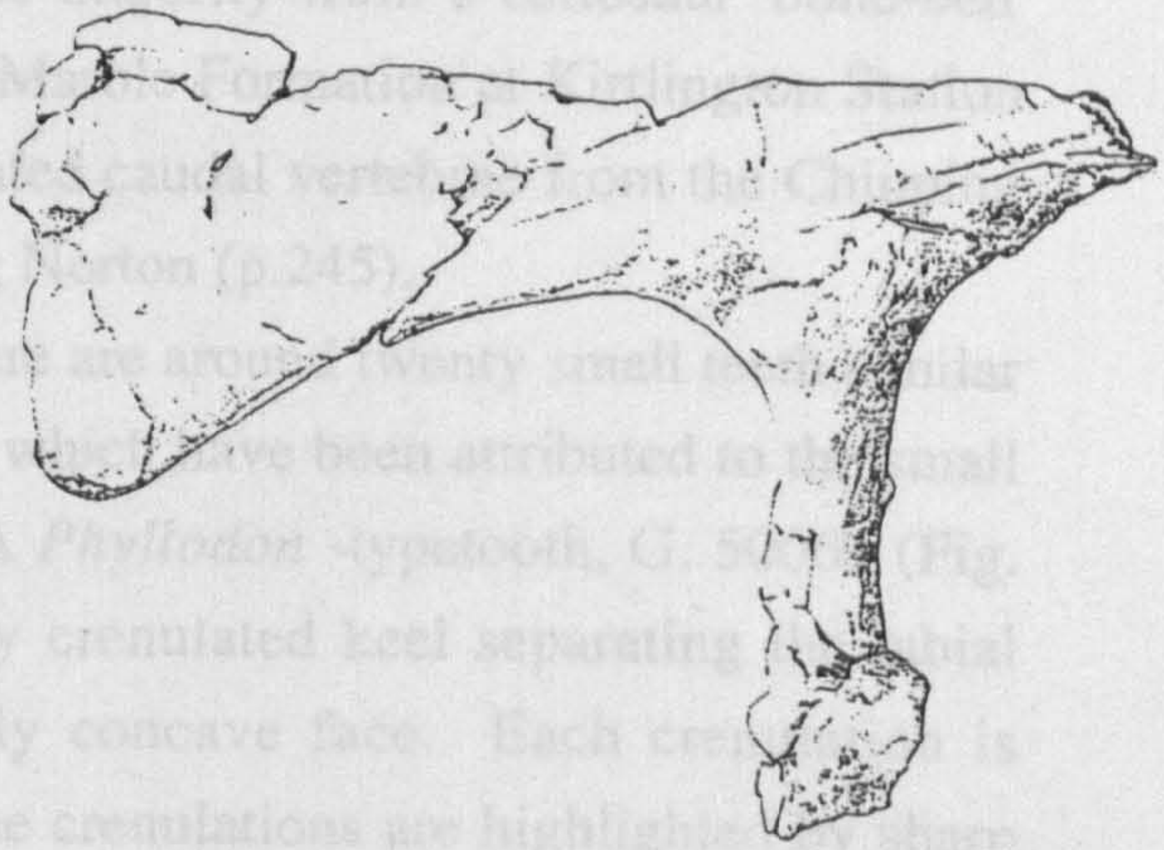

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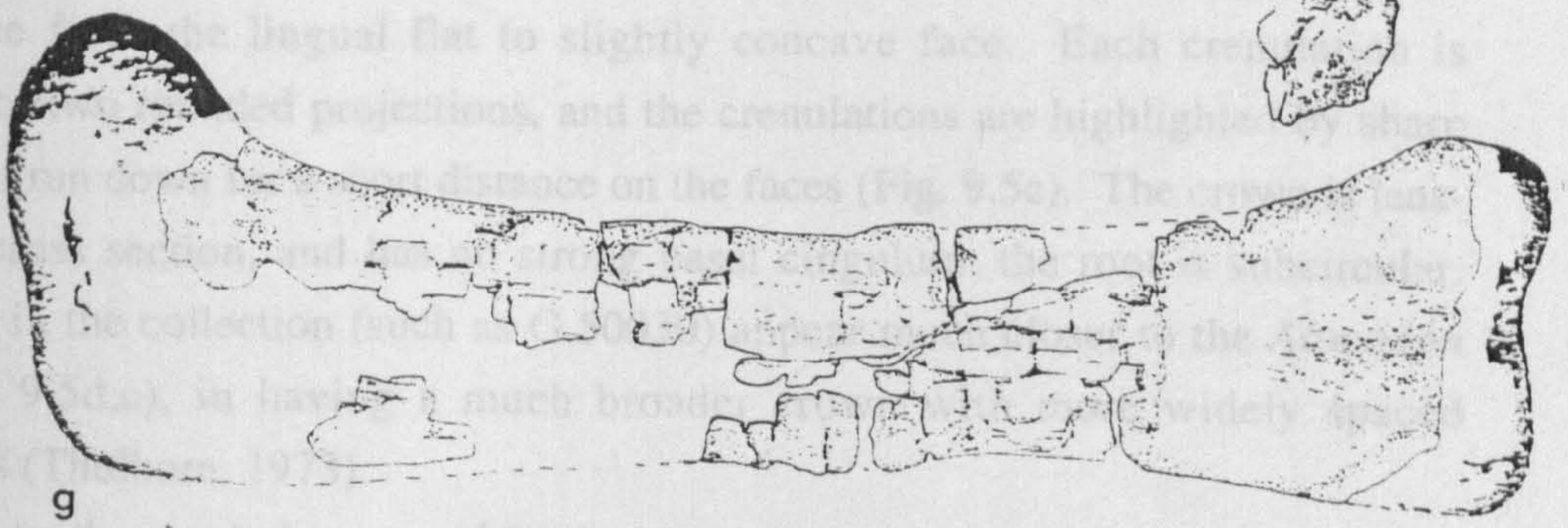

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G. 10014 and G. 10015 are fragments of the left ilium. A shattered, fragmentary femur, which would have measured 1.5m in length when complete (G. 10020; Fig. 9.4g), is the only identifiable limb bone, although undiagnostic shafts also occur.

The sauropod tooth crowns, G. 75712 and G.50051 (Fig. 9.4h-j) are concavo-convex, with a spatulate lingual face, and bear a slight off-centre ridge. The teeth are identified as sauropod by their characteristic spatulate shape, and lack of serrations along the margins.

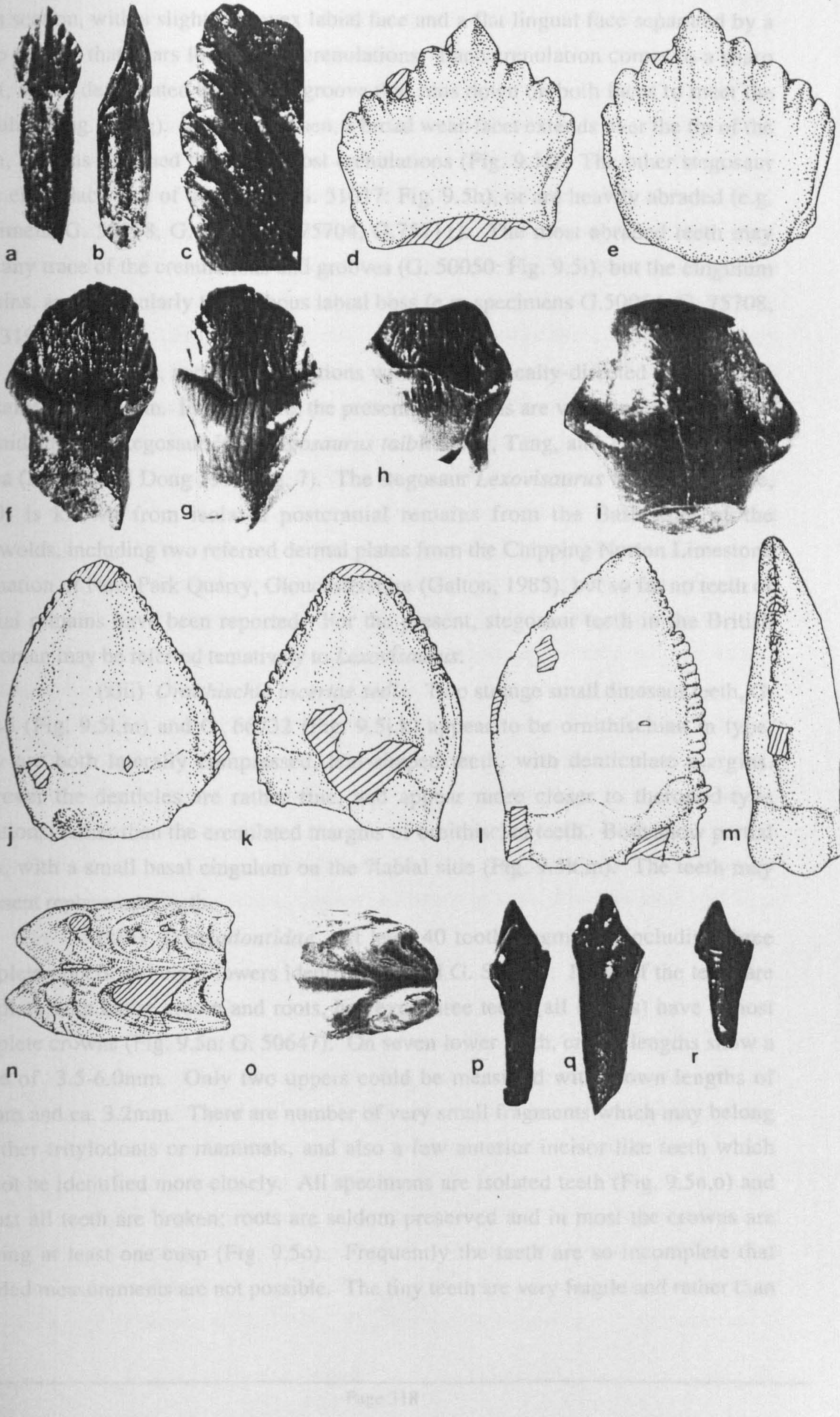
The postcranial elements and teeth resemble those of *Cetiosaurus oxoniensis* Phillips, 1871 in Oxford University Museum (OUM) and Leicester Museum (J. Martin, pers. comm, 1992). The Hornsleasow specimen represents a fairly large individual, about 25% larger than the Leicester cetiosaur (from the Oxford Clay), but comparable in size to the type specimen in OUM. *C. oxoniensis* is the typical sauropod taxon of the Cotswolds Bathonian, and many bones pertaining to this species have been recovered from the limestones of the Chipping Norton Formation (Owen, 1841, Phillips, 1871; section 9.1). The species was described by Phillips (1871) from several finds from the Bathonian of Oxfordshire, the majority from a cetiosaur 'bone-bed' (?three individuals: p. 250) within the Forest Marble Formation at Kirtlington Station Quarry, but also including the several associated caudal vertebrae from the Chipping Norton Formation of Chapel House, Chipping Norton (p.245).

(xi) *Hypsilophodontidae.* There are around twenty small teeth similar to the tooth genera *Alocodon* and *Phyllodon*, which have been attributed to the small ornithopod dinosaurs by Thulborn (1973). A *Phyllodon* -typetooth, G. 50069 (Fig. 9.5a,b) has a spatulate crown with a sharply crenulated keel separating the labial convex face from the lingual flat to slightly concave face. Each crenulation is composed of two rounded projections, and the crenulations are highlighted by sharp grooves that run down for a short distance on the faces (Fig. 9.5c). The crown is lens-shaped in cross section, and has no strong basal cingulum, the root is subcircular. Other teeth in the collection (such as G.50030) appear much closer to the *Alocodon* -type (Fig. 9.5d,e), in having a much broader crown with more widely spaced crenulations (Thulborn, 1973).

The teeth may belong to *Alocodon kuehnei* and *Phyllodon* sp. Thulborn, 1973, described on the basis of teeth from the Guimarota mine site, Portugal (Callovian-Oxfordian). Hypsilophodontids are known from the mid Jurassic, based on material of *Yandusaurus honheensis* He, 1979 from the Xiashaximiao Formation of China, although the group is more typical of the Cretaceous (Benton 1993b).

(xii) *Stegosauridae.* About 40 teeth, mostly broken and/or abraded are identified and described as stegosaurian. The most complete stegosaur tooth, G. 93101 (Fig. 9.5f,g) consists of a partial root and a crown. The characteristic broad bulbous smooth cingulum encompasses the middle part of the tooth and forms a bulbous boss

Figure 9.5. Hornsleasow indigenous vertebrate fauna: Ornithischian dinosaurs, mammal-like reptiles and mammals. Hypsilophodontids: (a-c) G. 50069, *Pyllodon*-type tooth with root, labial (a) and lateral (b) views (crown height: 2.1mm) and crenulated keel (c); (d-e) G. 50030, *Alocodon*-type shed crown, lingual (d) and labial (e) views (crown height: 1.4mm). Stegosaurs: (f-g) G. 93101, tooth with broken root and tip wear facet, lingual (f) and labial (g) views (height: 9.8mm including partial root); (h) G. 51017, broken crown with partial root, lingual view (height: 5.8mm including partial root); (i) G. 50050, worn shed crown, lingual view (crown height: 7.1mm). Unidentified ornithischians: (j-k) G. 66732, shed crown, labial (j) and lingual (k) views (crown height: 3.6mm); (l-m) G. 50045, tooth with broken root, labial (l) and lateral (m) views (height: 2.4mm including partial root). Tritylodont mammal-like reptiles: (n) G. 50647, *Stereognathus* complete lower crown, occlusal view (crown length: 5.7mm); (o) G. 51221, *Stereognathus* fragment of ?upper crown, occlusal view (length: 3.5mm). Mammals: (p-q) G. 50046, unidentified anterior premolar with single root, lateral (p) and lingual (q) views (height: 8.2mm including root); (r) G. 66570, unidentified incisor, lateral view (heigth: 5.3mm including root).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on the labial side of the tooth (Fig. 9.5g). The upper part of the crown is lens-like in cross section, with a slightly convex labial face and a flat lingual face separated by a sharp margin that bears four or five crenulations. Each crenulation comes to a sharp point, and is demarcated by a sharp groove that runs down on both faces to meet the cingulum (Fig. 9.5f,g). In this specimen, a broad wear-facet extends over the tip of the tooth, and has flattened the distal-most crenulations (Fig. 9.5f). The other stegosaur teeth either lack part of the crown (G. 51017: Fig. 9.5h), or are heavily abraded (e.g. specimens G. 50058, G. 93189, G.75704, G.75711). The most abraded teeth may lack any trace of the crenulations and grooves (G. 50050: Fig. 9.5i), but the cingulum remains, and particularly the bulbous labial boss (e.g. specimens G.50054, G. 75708, G. 93190).

The cingulum, and the crenulations with their vertically-directed grooves, are typically stegosaurian. Furthermore, the present specimens are very similar to teeth of the mid Jurassic stegosaur *Huayangosaurus taibii* Dong, Tang, and Zhou, 1982 from China (Sereno and Dong 1992, fig. 7). The stegosaur *Lexovisaurus vetustus* (Huene, 1910) is known from isolated postcranial remains from the Bathonian of the Cotswolds, including two referred dermal plates from the Chipping Norton Limestone Formation of New Park Quarry, Gloucestershire (Galton, 1985), but so far no teeth or cranial remains have been reported. For the present, stegosaur teeth in the British Bathonian may be referred tentatively to *Lexovisaurus*.

(xiii) *Ornithischia incertae sedis*. Two strange small dinosaur teeth, G. 50045 (Fig. 9.5l,m) and G. 66732 (Fig. 9.5j,k) appear to be ornithischian in type. They are both laterally compressed, leaf-shaped teeth, with denticulate margins. However the denticles are rather fine, and appear more closer to theropod-type serrations, rather than the crenulated margins of ornithiscian teeth. Both show partial roots, with a small basal cingulum on the ?labial side (Fig. 9.5k,m). The teeth may represent replacement teeth.

(xiii) *Tritylodontidae*. At least 40 tooth fragments, including three complete uppers and eight lowers identified by R.J.G. Savage. None of the teeth are complete with entire crown and roots, however three teeth (all lowers) have almost complete crowns (Fig. 9.5n: G. 50647). On seven lower teeth, crown lengths show a range of 3.5-6.0mm. Only two uppers could be measured with crown lengths of 2.8mm and ca. 3.2mm. There are number of very small fragments which may belong to either tritylodonts or mammals, and also a few anterior incisor like teeth which cannot be identified more closely. All specimens are isolated teeth (Fig. 9.5n,o) and almost all teeth are broken; roots are seldom preserved and in most the crowns are missing at least one cusp (Fig. 9.5o). Frequently the teeth are so incomplete that detailed measurements are not possible. The tiny teeth are very fragile and rather than

becoming abraded they will simply break in transport. The enamel on all teeth is stained black while the dentine and roots are dark brown-black in colour.

Although the material is very incomplete, R.J.G. Savage assigned the Hornsleasow material to the Bathonian genus *Stereognathus* (Benton *et al.*, 1995). *Stereognathus* is the only genus in the family known from Mid Jurassic times and there is nothing observable in the teeth to debar them from inclusion in *Stereognathus*. Only two species of *Stereognathus* are known: the type *S. ooliticus* based on two maxillary specimens from Stonesfield and *S. hebridicus* from Skye, based on an isolated upper molar and Dorset (Ensom, 1994). Waldman & Savage (1972) erected *S. hebridicus* on the basis of its larger size. While the size difference is marked, it may only indicate position in the jaw. In the Early Jurassic genus *Oligokyphus* , where associated dentitions are known, the cheek teeth vary in size along the jaw. The total number of *Stereognathus* teeth now known from Skye amounts to 27; most in range in length from 3.5 to 6.0 mm. The size of the isolated lower molar from the Forest Marble of Dorset is 5.3mm by 3.2mm (Ensom, 1994). From this it emerges that the Skye upper teeth are all bigger than any from Hornsleasow, but the Skye and Dorset lower teeth have a range that falls within the wide spectrum of the Hornsleasow specimens. At present it is impossible to resolve whether there are two distinct species or whether this represents the size range in the dentition of a single species. Consequently the best solution seems to be to designate the Hornsleasow specimens to one taxon, *Stereognathus* cf. *ooliticus*.

New tritylodontid material has been reported from a Forest Marble site at Kirtlington in Oxfordshire but is as yet undescribed (Freeman 1979). The family Tritylodontidae is distributed worldwide within a limited time range, Triassic to Middle Jurassic, with about a dozen genera known. *Stereognathus* occurs only in European Middle Jurassic sites and appears to mark the last appearance of the family. The type species from Stonesfield occurs alongside mammals as does *Stereognathus* on Skye and at Hornsleasow.

9.3.5. Mammalia

(i) *Triconodata or Symmetrodonta.* Two premolars (G.50017 and G 92402) described by R.J.G. Savage. The first has three cusps in a linear row, the middle cusp is the largest, and with one lateral smaller than the other. The second also a three cusped crown, with cusps in a linear row and near equal size, with the middle cusp being only slightly larger than the laterals.

(ii) *Symmetrodonta.* A right lower molar (G.50018) described by R.J.G. Savage as a very worn crown and tooth without roots. Also G.50020, a left lower molar and G.50039 a molariform tooth with three cusps arranged in a triangle (R.J.G. Savage in Benton *et al.*, 1995). The Middle Jurassic record of symmetrodonts

is extremely important as it was at this time that the symmetrodonts were giving rise to eupantotheres, the stock which were later to give rise to almost if not all later mammals. The eupantotheres are recorded from Middle Jurassic in Oxfordshire and Dorset (Freeman, 1979).

(iii) *Docodonta*. Only one tooth, G.92401, identified by R.J.G. Savage as an anterior lower premolar crown with one broken root.

(iv) *Multituberculata*. Two teeth assigned by R.J.G. Savage to this order. The first, G 60175, is a lower premolar crown with one root. Whilst the second, G 60172, is an upper first molar. This tooth is clearly a member of the family Paulchoffatiidae and very closely comparable with *Kuehneodon*. The genus is recorded in the Kimmeridgian site of Guimarota coal pit near Leiria in Portugal; three species are known, two of which are known only from mandibular rami and the third is known only from an upper jaw with partial dentition, which includes an M1. While this tooth has much similarity with the Hornsleasow tooth, another from Guirmarota is closer. As an isolated tooth the latter is only referred to as cf. *Kuehneodon.*

The Middle Jurassic is an important and little known period for mammal evolution, as it witnesses the very last survivors of the mammal-like reptiles with end members of the family Tritylodontidae and also the first appearance of the multituberculate mammals with members of the family Paulchoffatiidae. Both stocks were herbivorous and may have competed for food supplies. At least some of the multituberculates appear to have occupied similar ecological niches to the tritylodonts. and the multituberculates were themselves replaced taxonomically and probably ecologically in the Late Palaeocene/Early Eocene by the rodents.

(v) *Mammalia incertae sedis*. Five teeth are assigned to the mammalia *incertae sedis* by R.J.G. Savage (in Benton *et al.*, 1995) (Fig. 9.5p-r). One incisor, a canine and three anterior premolars (Fig. 9.5p-r) cannot be determined further with confidence.

Chapter Ten Vertebrate Taphonomy

10.1. Hornsleasow vertebrate assemblage parameters

The Hornsleasow vertebrate accumulation comprises a large and diverse assemblage preserved within a boggy paleosol. Vertebrate fossils are not usually abundant in paleosol profiles, and local concentration suggests preferential palaeoecological or preservational conditions for vertebrate remains (cf. Bown & Kraus, 1981). However, the Hornsleasow vertebrate assemblage is made up of disassociated and in many cases fragmented material, and has clearly undergone a complex taphonomic history prior to final exhumation. Therefore, in order to determine its unique derivation and taphonomic history, every specimen was assessed with equal care during the course of the study. This meant that all fragments no matter how indeterminate were as important in this investigation as an identified skeletal part (Metcalf, 1993).

There are two components to the Hornsleasow vertebrate assemblage, comprising the macrovertebrate remains of the *Cetiosaurus* dinosaur and the much larger mixed microvertebrate accumulation, which includes exotic and indigenous material (Chapter 9). The macrovertebrate remains are those specimens over a general size of 50mm in length (not including unidentifiable bone slivers), and range from fragments of large bones to complete vertebrae of the cetiosaur dinosaur. In this study only the largest pieces of identifiable bone (>50mm length), considered to be part of the dinosaur skeleton were studied. The total number of catalogued bones is 157; I studied 56 specimens (Metcalf, 1993). During the course of the taphonomic investigation I recorded the physical appearence, general association, size and description of the fossils within the Gloucester City Museum collections.

A similar study was also carried out for the microvertebrate remains. The microvertebrate material was recovered from the wet sieving of the clays, and is generally in the fraction size of 0.5mm - 20mm. The smallest fractions (<1-5mm, 45% of the sample) consist of tiny reptilian replacement teeth and jaws, some mammal teeth, flakes of bone, scute and fish scales (Metcalf, 1993). The most useful size category, in terms of taphonomic investigation, was the larger (5-20mm) and yields the most abundant remains (55% of the sample), consisting of fish scales and teeth, amphibian remains, reptile teeth, vertebrae, jaw and limb elements, scutes and bone fragments (Metcalf, 1993). The term "microvertebrate" is not used here in the biological sense (i.e. an animal under the net weight of one kilogram, *sensu* Behrensmeyer *et al.*, 1979), but as a reference to specimen size of under 20mm in total length (Metcalf, 1993). This

is not considered to be a significant problem in a taphonomic study, as for example, a 7mm tooth of a five tonne stegosaur dinosaur will behave taphonomically (when separated from the animal) more like the 5mm ilium of a discoglossid frog, than the same bone from the stegosaur.

Microvertebrate assemblages are not usually considered to be that useful in taphonomic and palaeoecological investigations (Maas, 1984; Lyman, 1994). They are considered to be derived by either fluvial sorting and concentration (Dodson, 1973; Wolff, 1973; Korth, 1979) or by accumulation of scatological material (Mellet, 1974). These processes are thought to create a biased assemblage by being too selective (i.e. predator preferences or hydraulic sorting). Pre-burial agents are most effective at destroying palaeoecological information to be derived from microvertebrate assemblages, but these are also produce the most easily resolved taphonomic delimiters on such accumulations (Maas, 1984).

In determining the taphonomic history of an assemblage of vertebrate remains the following parameters are considered to be the most useful:

(a) The structure of the accumulation

(b) The faunal composition

(c) The relative representation of skeletal elements in the fauna

(d) The condition of the specimens.

These parameters are described for both the macrovertebrate and microvertebrate assemblages separately in the following sections, and Figure 10.1. illustrates four specific categories to be considered in determining the taphonomic history of the Hornsleasow vertebrate assemblage. A typical taphonomy data sheet is given in Appendix F, listing the parameters for a box of 64 specimens of microvertebrates. Many of these specimens have been studied under the scanning electron microscope and photographed.

Many other taphonomic and palaeoecological principles have only been applied to mammalian assemblages based upon experimental work on modern bones and upon mammalian physiology (e.g. methods in Badgley, 1986; Behrensmeyer, 1975; Behrensmeyer & Dechant-Boaz, 1980; Korth, 1979). Their applicability to the lower vertebrates has not been tested, and therefore they may only have restricted value in assessing the palaeoecology of accumulations predominantly of fossil reptiles. These problems are discussed further in the following sections.

10.1.1. The structure of the accumulation

The structure of the vertebrate fossil accumulation essentially refers to the sedimentalogical and stratigraphic properties of the assemblage, and in most cases includes details which can only be recorded in the field. Firstly the association and articulation of the specimens within an assemblage should be determined. Articulation

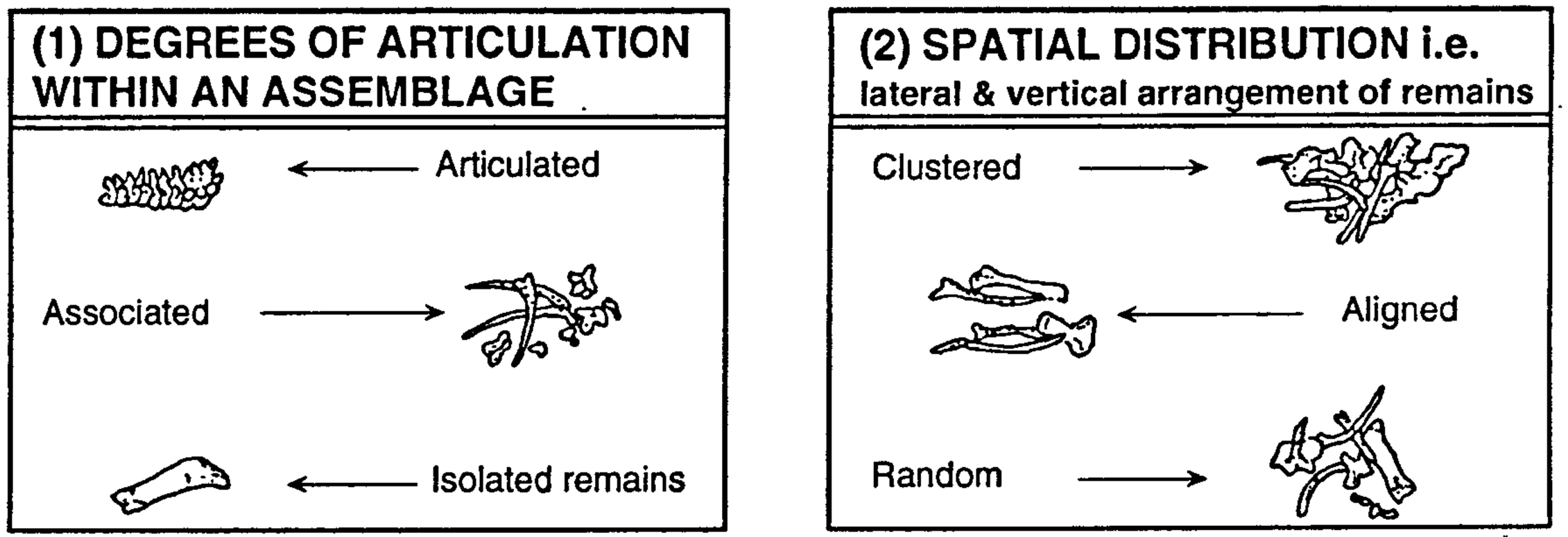


Figure 10.1. Generalised diagram of the four taphonomic parameters considered to be most useful in assessing the taphonomic history of a fossil vertebrate locality (after Badgley, 1986)

refers to the occurrence of skeletal elements in the same anatomical arrangement as in the living animal (Badgley, 1986). If the animal has become disarticulated, the skeletal parts may be still closely associated or completely disassociated and dispersed, occurring as isolated remains (Fig. 10.1). Disarticulation occurs after the animal dies and is caused by decay of soft parts, or the actions of predators and scavengers (Hill, 1979b). Further disassociation and dispersal is usually caused after the animal has been defleshed (although partial carcasses may be transported, as well as isolated bones and teeth), and includes hydrodynamic sorting by water (Voorhies, 1969; Behrensmeyer, 1975; Korth, 1979), indirect biological action (e.g. kicking and trampling by other animals; Andrews & Cook, 1985) and preferential destruction or preservation of bone material by non-biological modifiers (e.g. weathering, Behrensmeyer, 1975, Behrensmeyer & Dechant-Boaz, 1979). The presence of articulated skeletal material suggests little or no taphonomic modification, whilst that of closely associated remains indicates that the assemblage has not been secondarily sorted.

The macrovertebrate remains at Hornsleasow consist of a partial skeleton of the dinosaur *Cetiosaurus*, which although are in fairly close association over the restricted space of the excavation are totally disarticulated. Around 65% of the skeleton has been recovered, but it is impossible to determine whether the remainder lies under the unexcavated portion of the clay lens (Vaughan, 1989) or was removed by taphonomic sorting prior to burial. Also much of the material is badly fragmented and unidentifiable (Metcalf, 1993).

The microvertebrate remains are generally disarticulated and disassociated. However, probable clustering of bones and teeth is suggested by the recognition of several samples of similarly preserved specimens (e.g. those showing signs of acid digestion: section 10.2) of the same taxon occurring in close proximity in the sieved fraction (Metcalf, 1993). In this example, the presence of digested remains might indicate that the remains were derived from a disaggregated carnivoran coprolite (fossilised scat) or regurgitated material. Scatological accumulation of microvertebrate material is called "coprocoenosis" (Mellet, 1974).

Secondly, the lateral and vertical distribution of the fossils within each distinct sedimentary unit should be determined ('spatial distribution' of Badgley, 1986, p.332: Fig. 10.1). Badgley (1986) discovered that in a fluvial regime vertebrate material was scattered irregularly throughout the matrix in channel and crevasse-splay assemblages, whilst those from floodplain and channel margins were either randomly scattered or clustered. In most cases more than one taxon is represented in each cluster of bones, which suggests an outside agent of concentration, such as predator preferences (Maas, 1984) or trapping in a river system by hydraulic sorting (Dodson, 1973; Wolff, 1973; Korth, 1979). Alignment of vertebrate particles can also be assessed at this point, especially if palaeocurrent action is suspected. Such a task was undertaken upon the

large bones of the *Cetiosaurus* recorded upon Fig. 10.2, along with the large sections of wood found at the base of the karstic hollow. Although the wood particles showed varying, but possible palaeoflow alignment, the macrovertebrate material was randomly distributed about the floor of the hollow (section 5.1.2). The bones are scattered in one plane, lying flat upon the undulating surface of the limestone, or banked up against the sides of the hollow (Metcalf *et al.*, 1992). No bones were recovered in a near-vertical or vertical attitude (R. Vaughan pers. comm., 1991), which suggests that they were not trampled into the soft clay of the soil (Fiorillo, 1988a,b), or if trampling did occur (and there is evidence that it did, section 10.3.1), then it was with little soil cover upon the hard limestone karst. The bones are scattered over the whole of the base of the hollow, some arranged in clusters with associated pieces of fossilised timber (Fig. 10.2). The scattered distribution of the bones is thought to be derived from biological disintegration, and predator and/or scavenger activity upon the partial carcass of the *Cetiosaurus* prior to final burial (cf. Haynes, 1980, 1982; Hill, 1979; and section 10.2). The bones may then have formed a trapping device within the depositional low, for wood carried in by occasional flooding (Metcalf, 1993).

The 'stratigraphic span' (Badgley, 1986, p.333) of the accumulation refers to the variation in the vertical distribution of vertebrate fossil specimens within a sedimentary sequence. In this case, this parameter would refer to the arrangement of fossils within the two paleosol horizons, and is useful in determining time-averaging of faunas. The *Cetiosaurus* skeleton occurs only at the base of the lens, lying directly over the palaeokarstic limestone. Therefore its stratigraphic span is constrained and the assemblage has not been time-averaged. The fact that the skeleton shows signs of advanced weathering before burial in the paleosol, suggests a period of sub-aerial exposure (section 10.4.1).

The bulk sieving of the wet sediment does make it difficult to assess spatial distribution and stratigraphic range of microvertebrate remains within each individual clay horizon. Even without the problems of extraction mixing, the stratigraphic range would be difficult to determine for the two paleosol units. This is because as paleosol horizons the remains would be sufficiently churned up by pedogenic effects and bioturbation to destroy any sedimentary layering and time-average the vertebrate remains (Fiorillo, 1988b).

10.1.2. Faunal composition

The faunal composition of the Hornsleasow vertebrate assemblage was listed in Chapter 9. It appears that the macrovertebrate assemblage comprises the *in situ* remains of a single cetiosaur dinosaur. The microvertebrate accumulation is a mixed assemblage of fragmentary remains of small animals and small elements from larger beasts. Disregarding the cetiosaur skeleton, in terms of unadjusted specimen numbers, the most

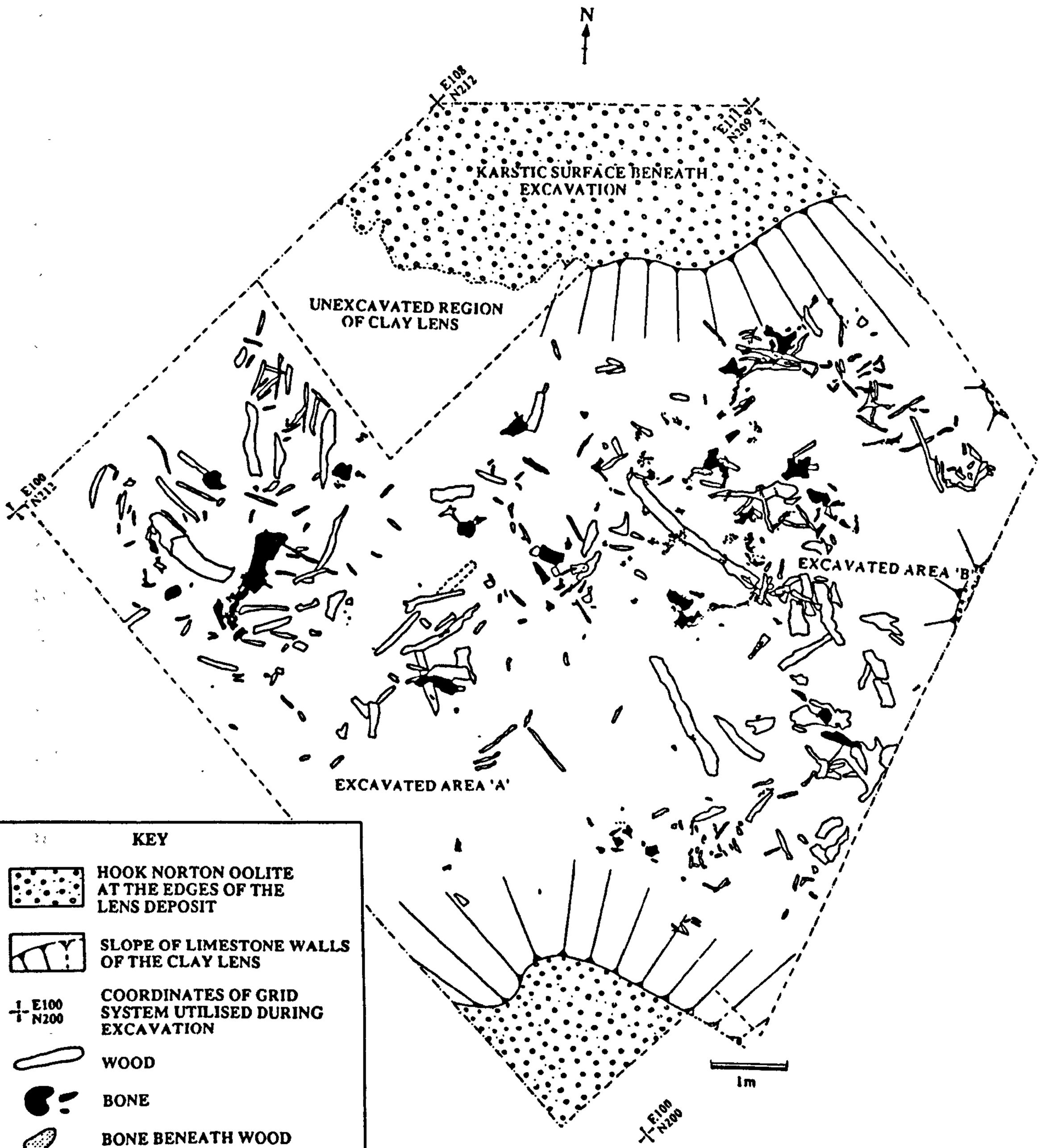


Figure 10.2. Sketch plan of the basal unit (karst and regolith layer) of the excavated site, showing distribution of wood and bone upon the gridded system (after Darlington, 1988, Metcalf *et al.*, 1992)

abundant animals in the Hornsleasow vertebrate assemblage are the crocodilians (53% of specimens), followed by fishes (14%), turtles (5%), pterosaurs (4%), mammal-like reptiles (3%), lepidosauromorphs (3%), champsosaurs (1%) and dinosaurs (1%). Amphibians and mammals represent less than 1% each of the preserved assemblage. These proportions are shown in Figure 10.3a,b. The proportions of specimens in the assemblage (Fig. 10.3b) suggest that reptiles make up over 70% of the identified remains, but this is largely due to an over abundance of their teeth (Fig. 10.4a; section 10.1.3). Therefore, the specimen abundance chart is not related to true ecological diversity. Although it is certain that the crocodiles and the turtles, both of which were represented by abundant post-cranial remains were a common component of the fauna and probably indigenous to the pool environment.

The rough proportions can be further developed in considering certain palaeoecological aspects of the faunal composition. For instance, the dominance of small aquatic vertebrates suggests a close proximity to freshwater for the assemblage and this is supportive of the sedimentary and invertebrate palaeontological evidence that the site represents a standing body of freshwater. This sort of taxonomic differentiation can provide much information upon the taphonomic nature of the assemblage as well. For instance, the predominance of microvertebrates of a certain size range or the numerical dominance of one or two species over all other taxa can indicate that an accumulation has been formed by a carnivore (Maas, 1985). Mammalian and avian predators are known to exhibit preferences for certain prey or have a specific size range (Pratt, 1989). The Hornsleasow assemblage is largely made up of an extinct reptilian community, hence any predator/prey activity can only be surmised. There is some evidence of predation on the bone assemblage (section 10.2), but it is difficult to assess its dominance upon the formation of the accumulation.

If the assemblage has been formed through fluvial or attritional processes then it is more likely to show a spread of faunal diversity, forming by chance rather than a function of prey-preferences of a carnivore (Pratt, 1989). A fluvial assemblage would also contain include a mixed fauna of terrestrial and aquatic animals (Behrensmeyer, 1975; Behrensmeyer & Hook, 1992). The spread of ecological niches amongst the fauna represented at Hornsleasow suggests that the assemblage formed by a combination of non-biological attritional processes (cf. Maas, 1984) and as a result of coprocoenosis (cf. Mellet, 1974). Some of the more exotic elements such as the teeth of large dinosaurs (e.g. the stegosaur teeth) may have been introduced into the hollow during depositional flooding episodes (Metcalf *et al.*, 1992).

Palaeoecological and diversity measurements such as the number of individuals per taxa, carnivore/herbivore ratios, and proportion of different age groups, and the sexes within an accumulation (Badgley, 1986; Pratt, 1989) are much more difficult to assess for the microvertebrate accumulation. The ability to identify most of the remains

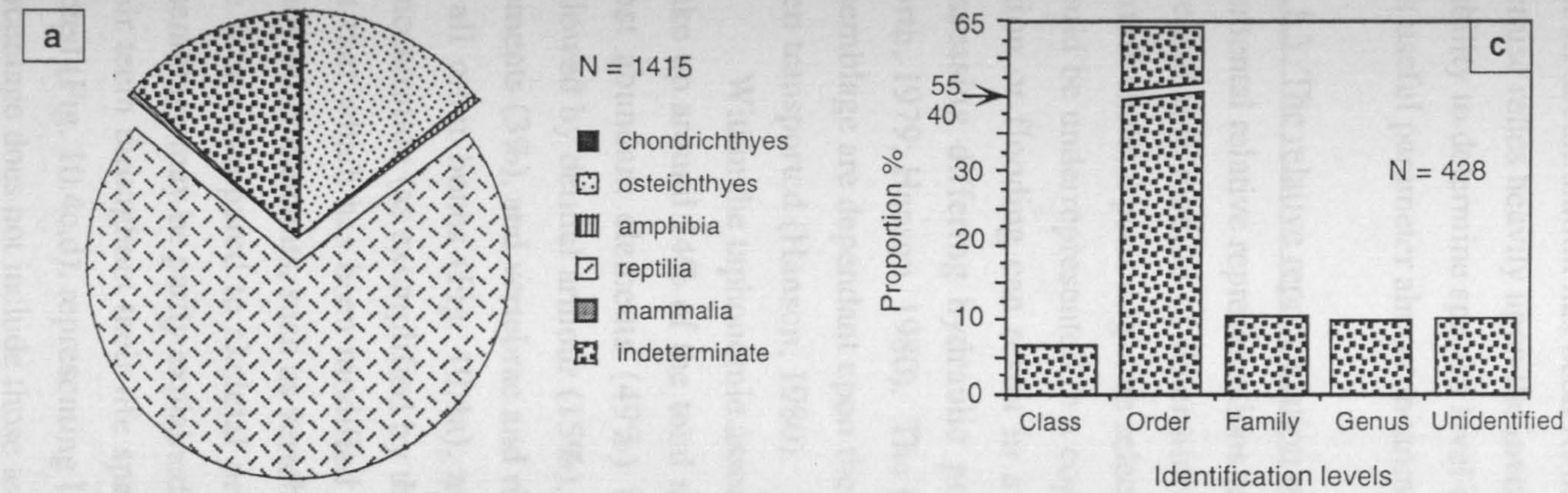


Figure 10.3. The faunal composition of the microvertebrate assemblage (a-b). The fauna is dominated by reptiles at a coarse taxonomic level (a), which in turn are dominated by crocodile remains (b), which make up more than 55% of the vertebrate assemblage. (c) shows the relative degrees of identification possible for the microvertebrate remains.

to only ordinal level (Fig. 10.3c) renders any possible change in faunal diversity between the two clay units extremely difficult to determine as both paleosols contain the same families and genera. However, those species which have been identified do not appear to change across the two horizons, although some species are less common in the Green Clay. This is probably a reflection of palaeoenvironmental change though as the aquatic species are those most affected.

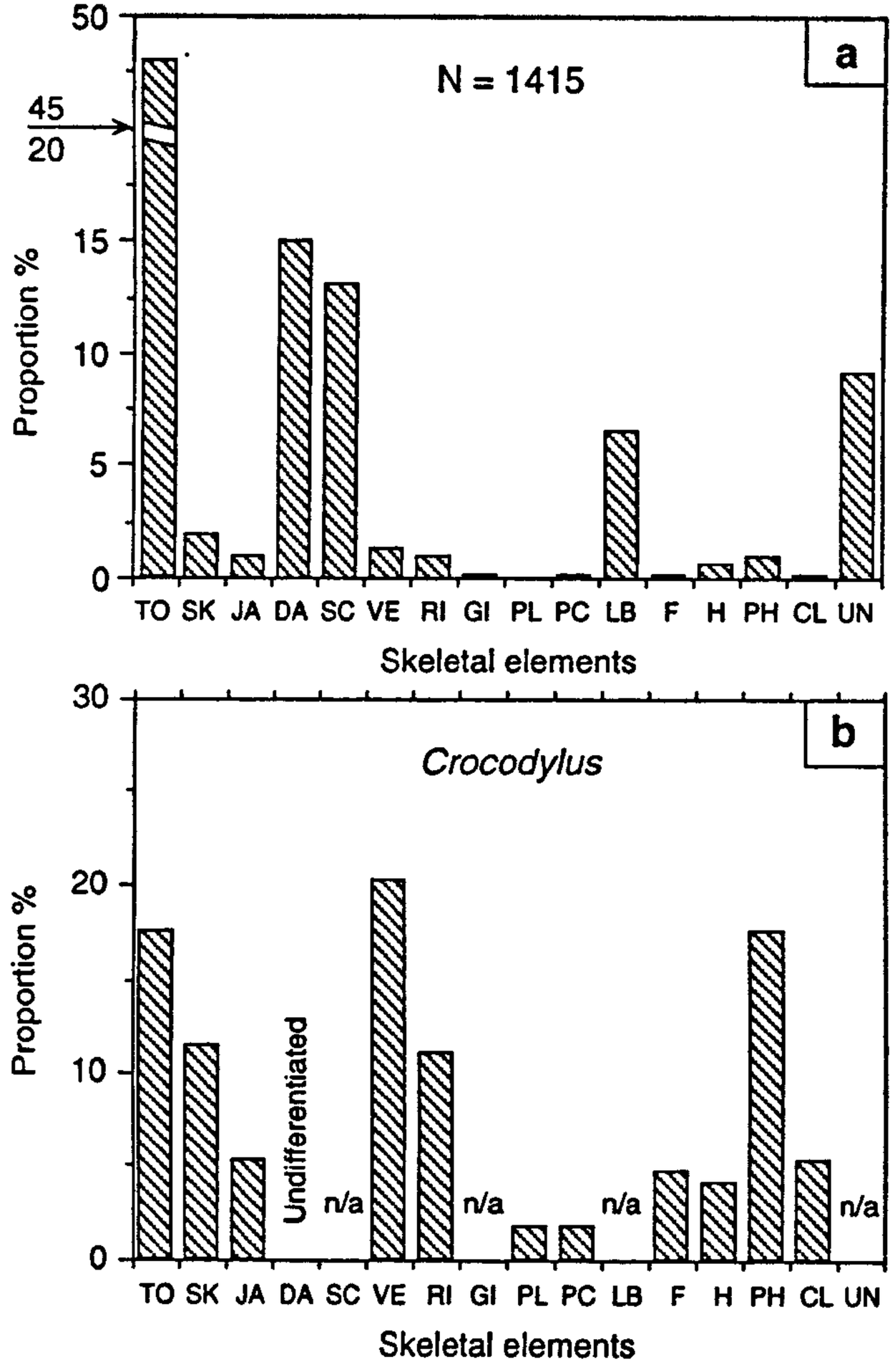
In mammalian assemblages a high proportion of juvenile remains might indicate that an accumulation has been formed from predator activity (Badgley, 1986), but this premise relies heavily upon the investigator's ability to identify juvenile specimens. The inability to determine species level classification in the Hornsleasow assemblage makes this useful parameter almost meaningless.

10.1.3. The relative representation of the skeletal elements

Elemental relative representations (as defined by Wolff, 1973 and Korth, 1979) provide a reasonable method of determining the source of a microvertebrate accumulation. Certain skeletal parts might be selectively discarded or destroyed by predator action and would be underrepresented in a coprocoenosis. Also hydrodynamic sorting by fluvial action or flooding can result in a winnowed deposit, with various different bones possessing differing hydraulic properties (Voorhies, 1969; Behrensmeyer, 1975; Korth, 1979; Hanson, 1980). The types and numbers of bones represented in such an assemblage are dependant upon the current speed and the distance that the element has been transported (Hanson, 1980).

Within the taphonomic assemblage at Hornsleasow unidentified bone fragments make up around 14% of the total microvertebrate sample (Fig. 10.4a). Teeth are the most abundant elements (49%) in the microvertebrate assemblage (Fig. 10.4a), followed by dermal armour (15%), fish scales (13%), limb bones (9%), skull and jaw elements (3%), and vertebrae and ribs (2%). Teeth occur in a disproportionate number to all other bones (Fig. 10.4a), and this assemblage does not resemble a skeletal concentration (as exemplified by the skeletal proportions of the crocodile: Fig. 10.4b) but that which has been modified by post-mortem modifications and sorting. Also many other elements, such as vertebrae and phalanges, are severely underrepresented in the fauna compared to skeletal proportions. The abundance of teeth in a reptilian assemblage may be partly explained by the fact that reptiles continually shed and replace their teeth throughout their life span. Less than 20% of the teeth at Hornsleasow are rooted (Fig. 10.4c,d), representing loss from the jaws of dead individuals (note that this percentage does not include those teeth still attached to jaw fragments).

As the accumulation does not represent a skeletal assemblage, then the role of the taphonomist is to determine why this is so, for instance are the elements underrepresented in the fauna because of severe fragmentation of the specimens making



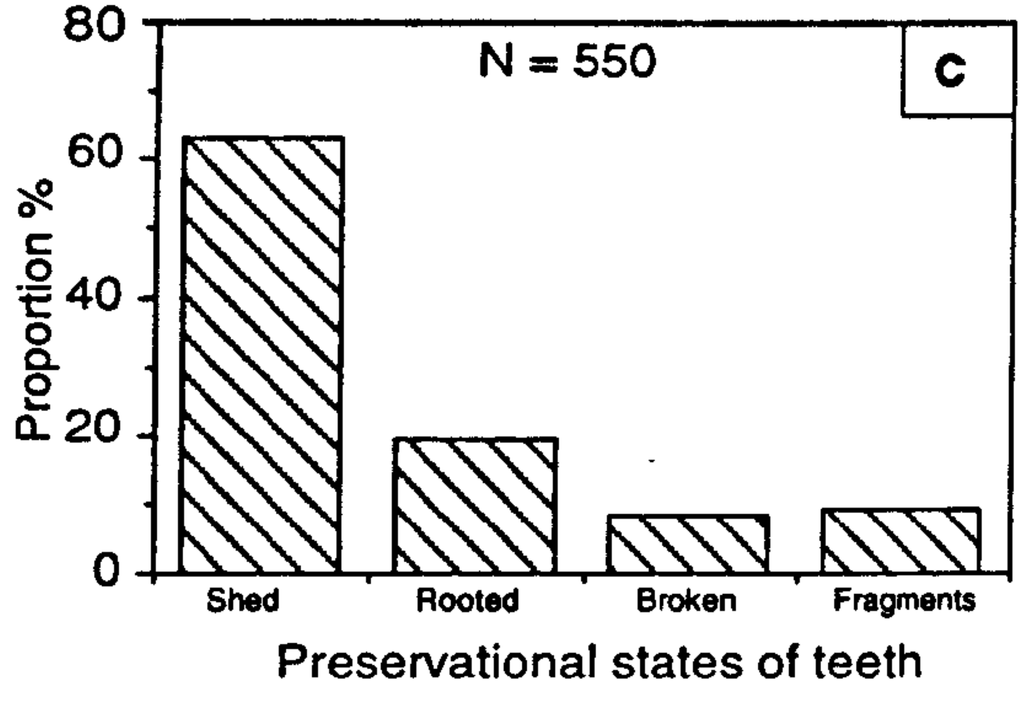


Figure 10.4. The relative representation of skeletal elements in the microvertebrate accumulation (a) and a modern reptile (*Crocodylus*) (b). Abbreviations for skeletal elements: 'TO' teeth; 'SK' skull parts; 'JA' jaw material; 'DA' dermal armour; 'SC' fish scales; 'VE' vertebrae; 'RI' ribs; 'GI' girdle material undifferentiated; 'PL' pelvic bones; 'PC' pectoral bones; 'LB' limb bones undifferentiated; 'F' fore-limb bones; 'H' hind-limb bones; 'PH' phalanges; 'CL' claws; 'UN' unidentified bone; 'DE' unidentified dermal bone. The most abundant fraction within the assemblage are the isolated reptile teeth (c), with Nt = 47%. Of this proportion over half (Ns = 63%) are isolated shed crowns (d). The number (N) of specimens is indicated.

identification almost impossible? This could be the solution to the low number of identified limb bones (Fig. 10.4a), as many occur as broken shafts or epitheses, making identification extremely difficult. The second reason for the impoverished skeletal assemblage could be that it has suffered some sort of taphonomic sorting prior to final burial. This can be determined by comparing the accumulation to those within equivalent bone accumulations seen in natural and simulated environments of deposition (Behrensmeyer, 1975; Behrensmeyer *et al.*, 1979; Behrensmeyer & Dechant-Boaz, 1980; Bown & Kraus, 1981; Dodson, 1973; Voorhies, 1969) and those recovered from scatological accumulations (Dodson, 1973; Maas, 1985; Mellet, 1974; Korth, 1979). The idea that the elemental assemblage might have been sorted is discussed further in the following sections.

10.1.4. The general condition of the specimens

A detailed study of the surface condition of fossil bone can provide useful information as to which processes acted upon the material prior to and during fossilisation (Fiorillo, 1988a). Studies of bone-surface modifications have largely been upon modern bone assemblages (e.g. Mayhew, 1977; Behrensmeyer, 1975, 1978; Korth, 1979; Haynes, 1980a,b, 1982). Detailed studies of taphonomic alteration of surface features of fossil bones are much more rare (e.g. Hunt, 1978; Shipman, 1981; Fiorillo, 1988a). The assessment of "bone damage" includes several easily distinguishable modifying processes, such as physical modification and chemical destruction of bone material. These are considered very briefly below and at greater lengths in the following relevant sections documenting the taphonomic history of the assemblage.

Observations were made on the surface condition of a representative sample of the macrovertebrate and microvertebrate remains from Hornsleasow Quarry. All macrovertebrate material was studied under laboratory conditions with lighting from above, and was studied for at least 20 minutes. The microvertebrate sample comprised taxonomically unsorted material in boxes (i.e. that sieved at Gloucester City Museum). This was studied under a conventional microscope well lit from the sides and above, and some specimens were also inspected under the SEM. Bones were excluded from this study only if the external surface had been removed, although efforts were made to understand how this may have happened. Several categories of surface feature and bone modification were examined in this study, and these were : (1) fracture patterns; (2) presence or absence of bite marks or gnaw marks; (3) occurrence and abundance of trample marks; (4) weathering stage; (5) abrasion index; (6) compaction damage; (7) occurrence or absence of mineralisation. These categories are listed on Table F.1, Appendix F.

The best preserved in the microvertebrate assemblage are teeth and jaws, and of the post-cranial remains the most complete are either the smallest bones or those with

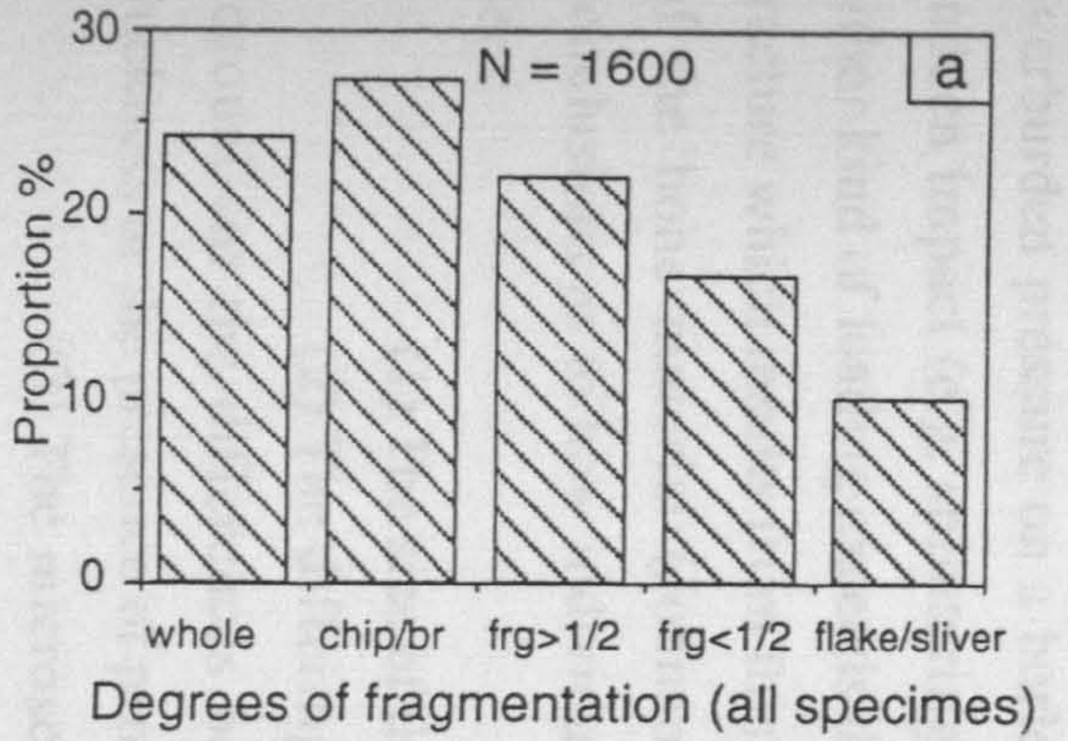
high densities or a compact shape (e.g. vertebra centra, and bones of the hands and feet. A similar result was recorded from the micromammalian assemblage recovered from boggy paleosols in the Willwood Formation (Eocene) by Bown and Kraus (1981). Virtually all of the microvertebrate remains are fragmentary (Fig. 10.5a), although most specimens are only chipped or lacking less than half their mass (Fig. 10.5a). Most of the compact and dense elements, such as teeth, are well preserved and only show chipping or partial fragmentation (Fig. 10.5b-d). More delicate remains such as vertebrae have lost processes, whilst limb bones and jaw fragments are incomplete and have been broken at their articular ends (Fig. 10.5f,g,i,k). Breakages tend to be both sharp and unabraded (Fig. 10.5d,e,i,k) or rounded and worn (Fig. 10.5b,c,f) depending on the post-fragmentation taphonomic history.

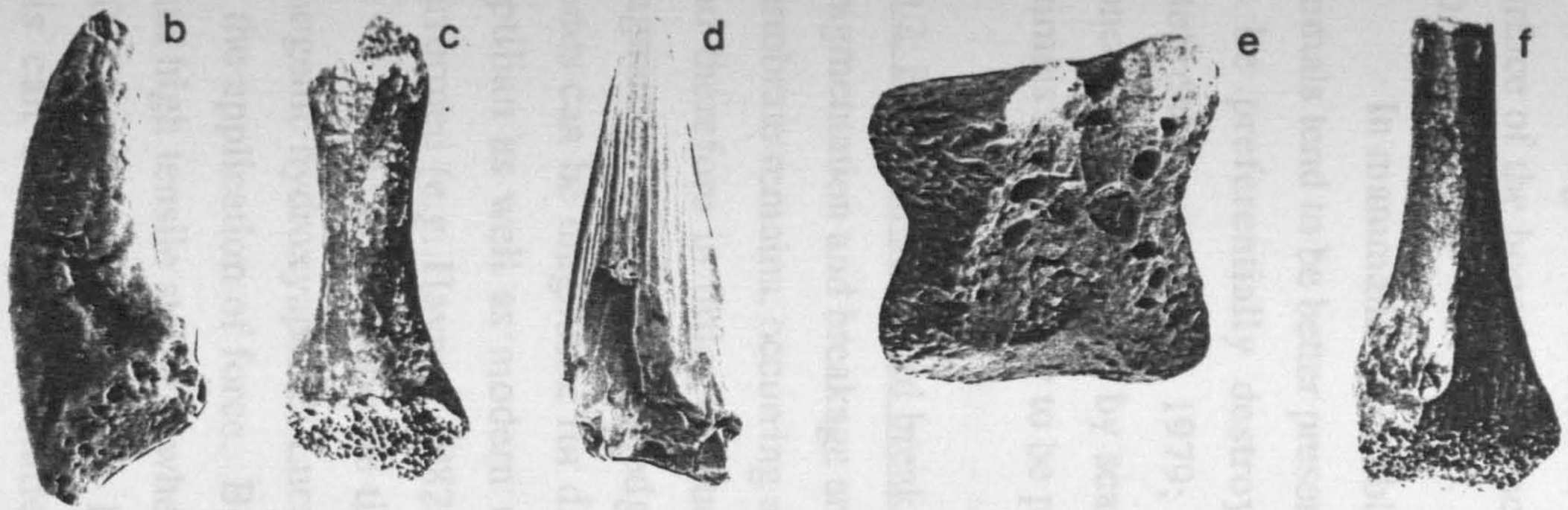
The three types of fracture pattern defined by Myers *et al.* (1980) were present in the Hornsleasow vertebrate assemblage. The first is the 'spiral' or 'green-stick' fracture common in fresh-bones, and defined by Myers *et al.* (1980) as a breakage occurring in a bone at an angle other than 90° to the direction of the bone fabric. This type of fracture is indicative of a bone which has broken prior to any type of alteration to the bone material. The second fracture pattern is that which occurs at right angles to the fabric of the bone, this is called a 'transverse' fracture (Myers *et al.*, 1980). This type of fracture usually occurs some time after alteration of the original bone has taken place (Fiorillo, 1988a). The third type of breakage forms parallel to the bone fibre, and is known as 'longitudinal' cracking (Myers *et al.*, 1980). This type of breakage results from desiccation of the bone material prior to fossilisation. Fracture patterns are discussed in more detail in section 10.2.

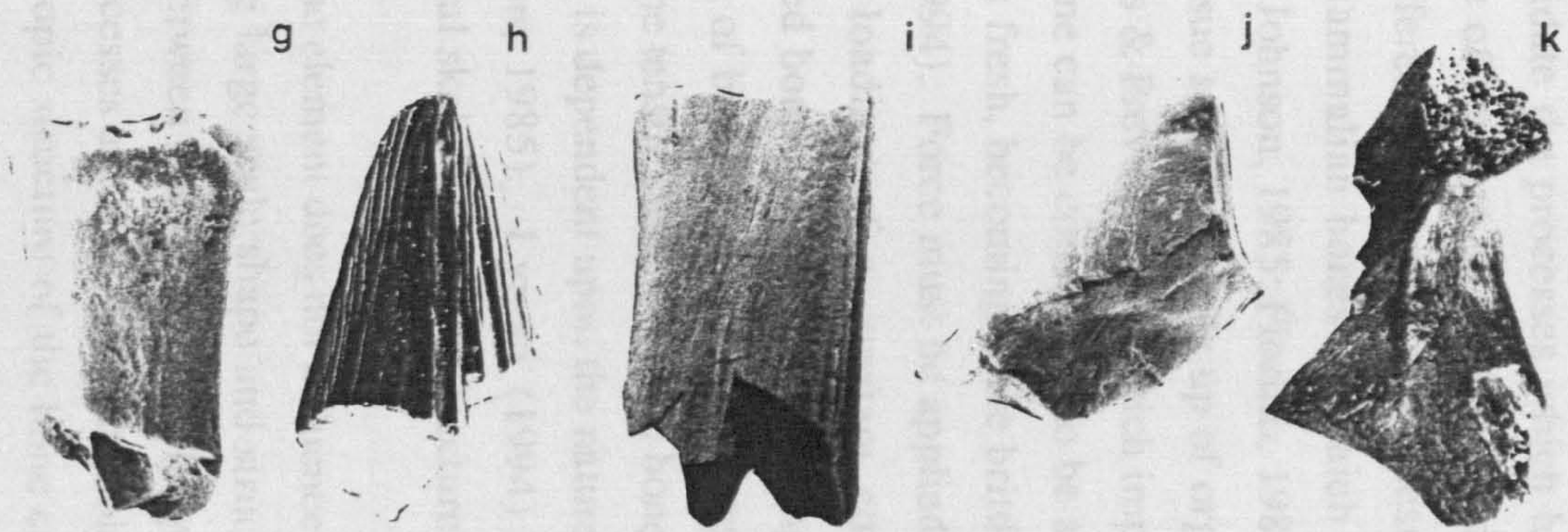
Most of the bones in the microvertebrate assemblage are relatively unweathered (section 10.5). Bite marks are also extremely rare, although some of the bones may have been broken by predator or scavenger action (section 10.3) and bone-bearing coprolites are also fairly common, suggesting biological activity. A few teeth have lost virtually all their enamel, but are relatively unbroken or unabraded, suggesting that they have passed through the digestive tract of a predator (section 10.3).

Some of the bones and teeth show root acid attack upon their surfaces (section 10.4.2) and much of the material has been broken during burial and compaction. Early diagenetic permineralisation of bones and teeth seems to have occurred, and there is no positive evidence that material was destroyed or preferentially preserved by differing chemical conditions in various parts of the soil unit. Much of the material contains limonite infills or crusts (section 10.7).

The macrovertebrate remains are fairly weathered upon the surface of the bone facing upwards (section 10.5) and broken surfaces appear to have suffered from a degree of abrasion (section 10.6). Primary biological traces such as trample marks and bite marks are not common (sections 10.3 & 10.4), but these may have been obliterated







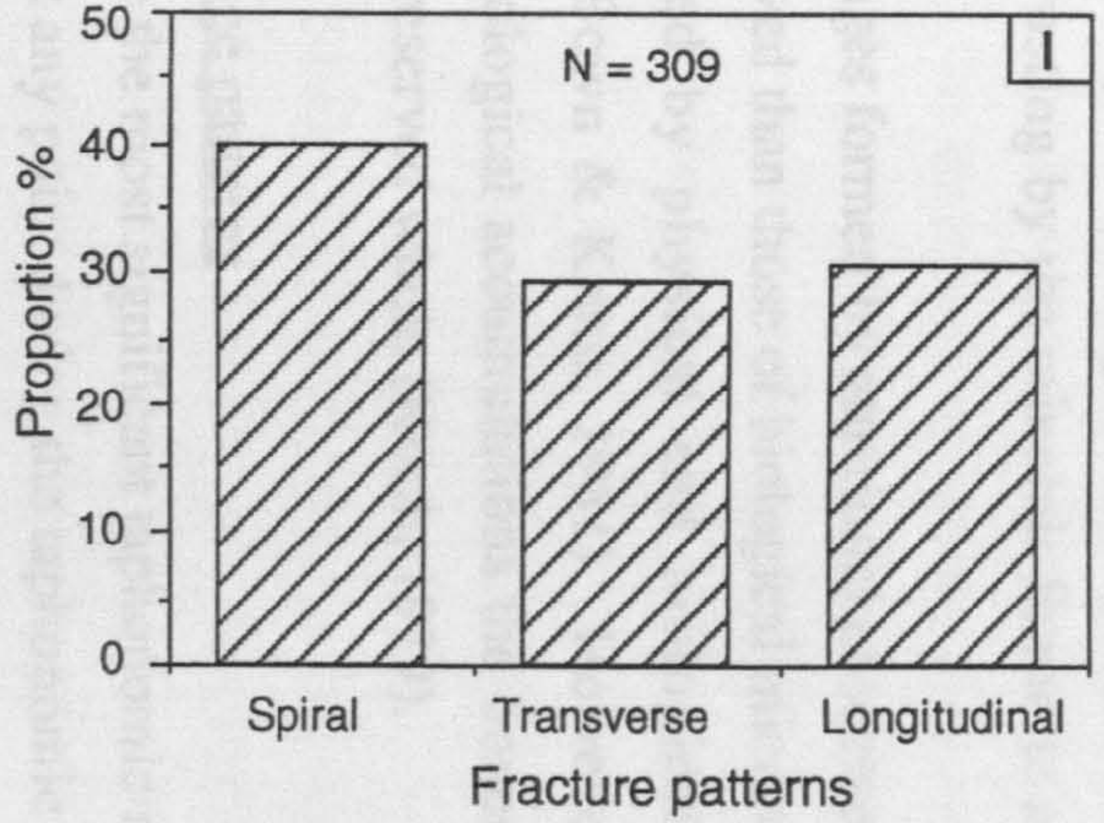


Figure 10.5. The majority of the microvertebrate specimens show a degree of fragmentation (a) ranging from chipping and minor breakages ('chip/br') on otherwise whole specimens (b-c), to fragments ('frg') over half (d-f) and less than half (g-h) of the specimen, and slivers of bone or tooth (i-k). Details of specimens (b) G. 52143, reptile claw (length = 8.5mm); (c) G. 71111, limb element showing damage to articular ends (length = 18mm); (d) G. 61405, splintered crocodile tooth (length = 1.7mm); (e) G. 51241, broken fish vertebra (3.4mm across); (f) G. 71112, limb missing articular end (length = 12.8mm); (g) G. 61613, limb shaft fragment (length = 1.5mm); (h) G. 56905, fragment of crocodile tooth (length = 1.6mm); (i) G. 57724, splintered limb shaft (length = 4mm); (j) G. 61404, crown fragment of tritylodont tooth (1.4mm basal width); (k) G. 71113, bone fragment (15.5mm across); (l) Fracture patterns of broken microvertebrate material in the assemblage.

by the considerable weathering and mineral growth. The bones are quite broken and much of this appears to have taken place after burial, compaction shears and large-scale vein calcite growth is seen in many of the larger bones (section 10.7). The weathered surface of the bones show encrusting by the minerals limonite and pyrolusite (section 10.7).

In mammalian assemblages formed by attritional processes the remains of larger animals tend to be better preserved than those of biological microvertebrates, which tend to be preferentially destroyed by physical and biological taphonomic agents (Behrensmeyer *et al.*, 1979; Bown & Kraus, 1981). However, in microvertebrate concentrations formed by scatological accumulations the bones and teeth of smaller animals are more likely to be preserved whole (Korth, 1979).

10.2. Fragmentation and breakage patterns

Fragmentation and breakage are the most significant taphonomic modifications affecting vertebrate remains, occurring at any point during the taphonomic history of the material and therefore in order to elucidate the processes which have modified a particular fragmented bone, a knowledge of fracture patterns is needed. Breakage patterns of bones can be diagnostic for different fragmentation processes, and this holds true for reptilian as well as modern mammalian bones on which the original studies were performed (e.g. Haynes, 1982; Johnson, 1985; Fiorillo, 1988).

Bone is a composite tissue structure made up of organic and mineral material (biogenic hydroxyapatite: Lucas & Prevot, 1991), which impart to it an unique reaction to the application of force. Bone can be considered to be an almost elastic substance with high tensile strength when fresh, becoming more brittle when dry or mineralised (after fossilisation) (Lyman, 1994). Force must be applied to a bone to break it, and this can be from the 'static loading' of a constant compressive pressure (e.g. overburden pressure on a buried bone) or 'dynamic loading' involving the force of a sudden impact (e.g. splintering of bone in a predators' jaws) (Johnson, 1985). When either kind of loading exceeds the tensile strength of the bone, it will break. The type of fracture which results from this is dependent upon the nature, shape and microstructure of the bone material (Johnson, 1985). Lyman (1994) provided several general conclusions as to how individual skeletal elements fracture and these are summarised here:

(1) The size of the element does not influence bone fracture patterns.

(2) The differing large scale shape and structure of bone elements will account for the differences between fracture location upon them. Differences in thickness or the presence of processes are usually responsible for this phenomenon.

(3) The microscopic structure of the bone can also exert some control upon the fracture form and pattern.

(4) Lightly weathered bone usually fractures obliquely, but heavily weathered bone tends to fracture transverse to the grain of the bone. Heavily weathered bones fracture more easily with less force than fresh or lightly weathered bone.

(5) Static loading of a bone causes fractures to occur at a point opposite to the point of loading, whereas dynamic loading enables several fracture points to be established radiating away from the point of impact.

(6) Spiral fracturing is more common than other forms of fracture pattern under static and torsional loading, although it is less common when dynamic forces are applied to fresh bone.

(7) Dynamic loading results in more rounded fracture ends than statically fractured bones.

As outlined in the preceding section, fresh bone breaks differently to dry or mineralised bone. The different types of fracture are summarised in Fig. 10.6 (after Shipman *et al.* 1981 and Marshall, 1989). Fresh or 'green' bone is made up of an organic and mineral phase and is mechanically quite strong behaving as a part elastic - part inflexible substance (termed 'visco-elastic': Johnson, 1985). Experimentally broken fresh bones fracture through lines of weakness producing splintered, saw-toothed, oblique or curved fractures ('spiral' fractures: Fig. 10.6b,c,d,g). Typical spiral fractures occur throughout the main body of the bone element, at an angle to the fabric, however, spiral fractures may be irregular, curved, splintered or jagged corresponding to the force applied to the element (Haynes, 1982; Fiorillo, 1988a). They may also be flat and smooth diagonal fractures at a low angle to the long axis of elements which have torsional stresses applied to them in life (e.g. the humerus; Hill, 1980: Fig. 10.6e).

When dry or mineralised, a bone behaves in a brittle fashion, under static loading it is thought to be stronger than fresh bone, but fails under lower dynamic forces, as its shearing strength has been reduced (Lyman, 1994). The resulting fractures usually occur transverse to the long axis of the bone under stress, and form major, clean, planar breakages (Fiorillo, 1988a: Fig.10.6f). Such fractures occurring in fossilised bones are called 'transverse fractures' and are probably caused by the response of the fossilised tissues to diagenetic pressure in a similar manner to the cleavage planes of the enclosing rock (Shipman, 1981). Flakes of bone can be removed when the bone is dried or mineralised, to produce characteristic chipped, columnar or stepped breakages and rectangular bone fragments (Shipman, 1981: Fig. 10.6a,d). Bone flaking can also occur with severe longitudinal cracking caused by weathering (Fig. 10.6h) and in some cases by predator chewing. Flaking occurs when a mosaic or tessellated pattern of minute surface cracks appears during desiccation of the bone. Columnar breakage patterns are often thought to be more diagnostic of dried rather than mineralised bones, as they seem to be associated with splitting along longitudinal cracks

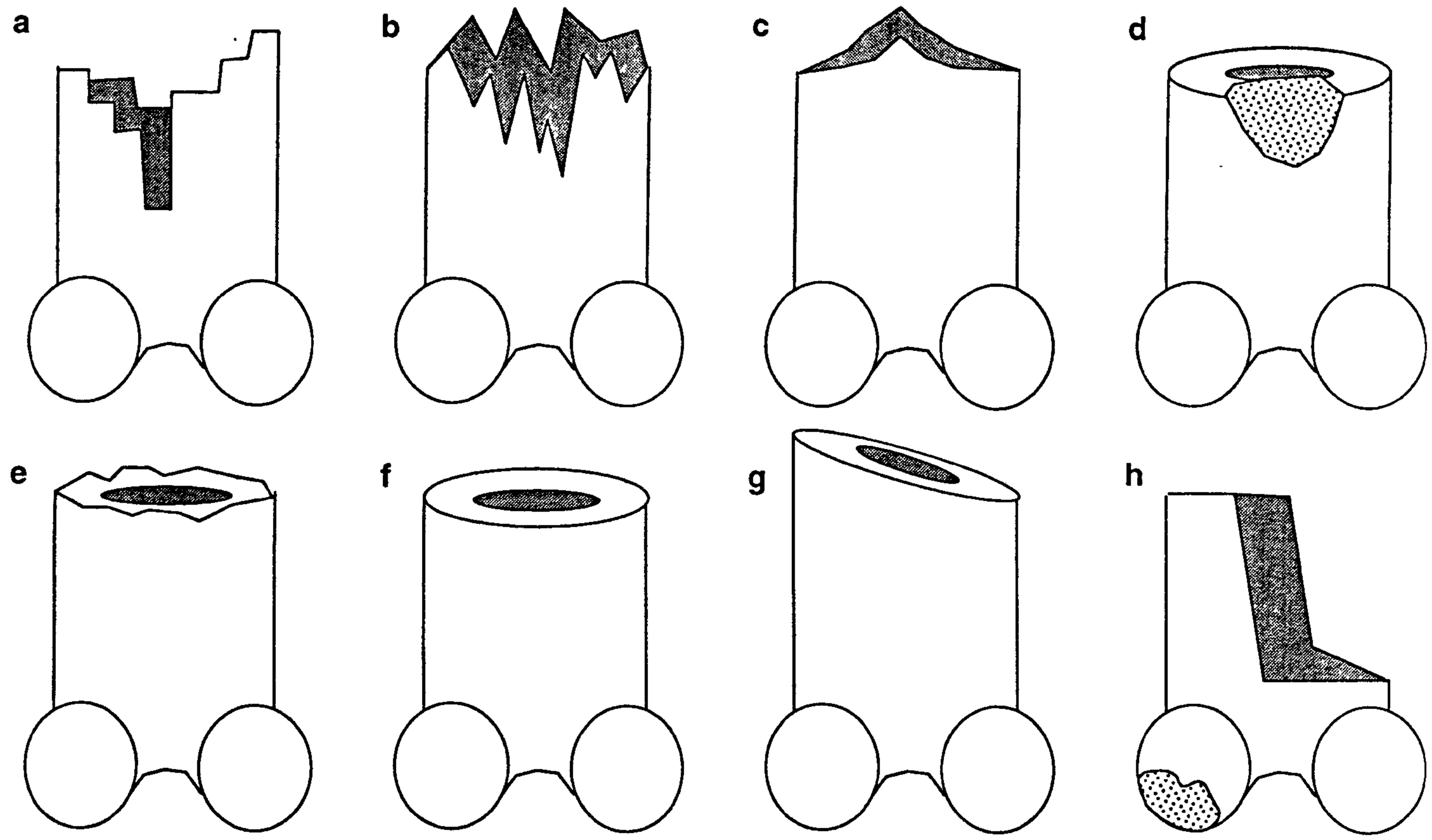


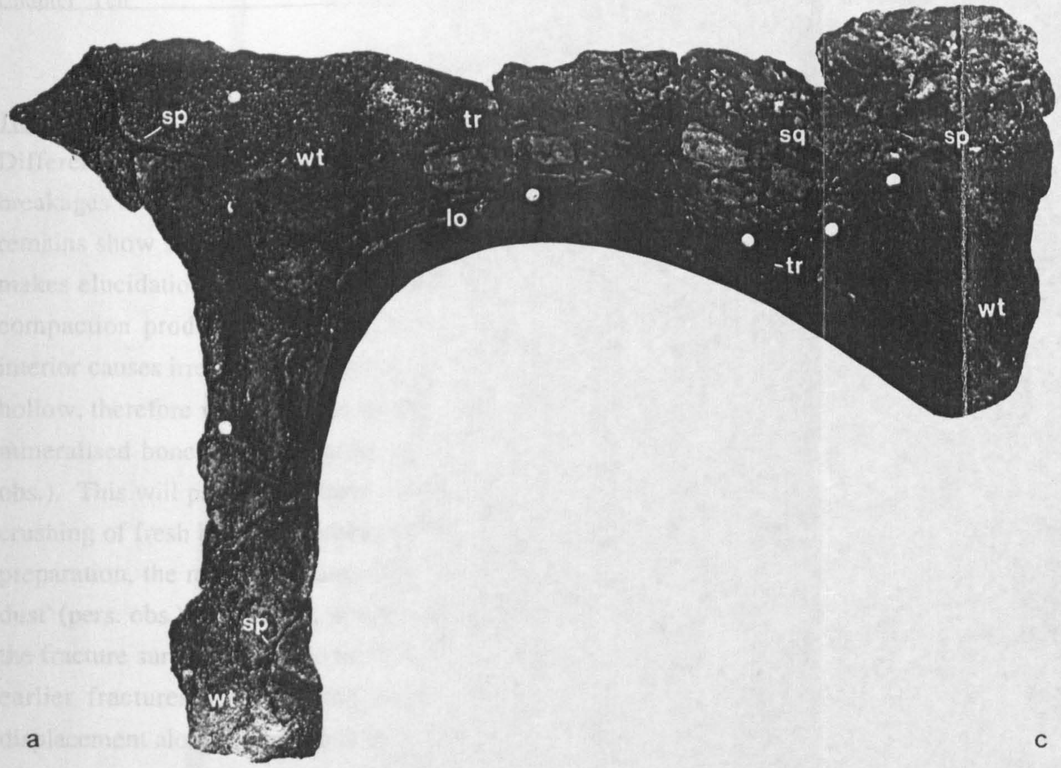
Figure 10.6. Fracture types (after Shipman *et al.*, 1981 and Marshall, 1989), as shown upon a generalised limb bone. (a) a stepped or columnar fracture of the shaft; (b) a sawtooth fracture of the shaft; (c) a V-shaped breakage through the shaft; (d) flaking of the outer layer of the bone; (e) an irregular fracture perpendicular to the shaft; (f) a smooth transverse fracture through the bone; (g) a spiral or oblique fracture through the shaft; (h) longitudinal breakage through the bone and irregular breakage of the articular end, exposing cancellous tissue.

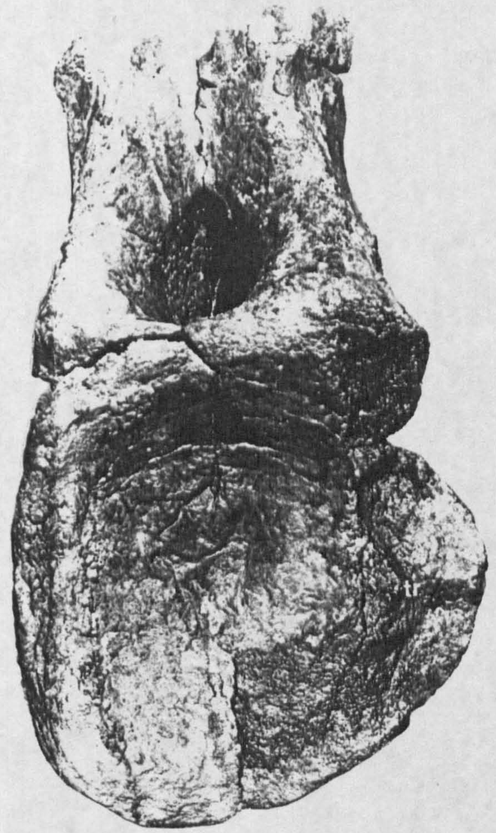
caused by desiccation. Longitudinal cracks are the most easily distinguished desiccation fractures as they run parallel to the grain of the bone or tooth, and occur as both large-scale and microscopic features, these are often attributed to surface weathering, but can also occur in buried and sub-fossilised bone material due to minute physiochemical differences in the surrounding conditions (C. Trueman, pers. comm., 1995). Bone chipping is usually caused by movement of the element, such as the battering encountered in a fluvial regime or by kicking and trampling by other animals (Andrews & Cook, 1985). Chipping of fossilised bone causes conchoidal fractures.

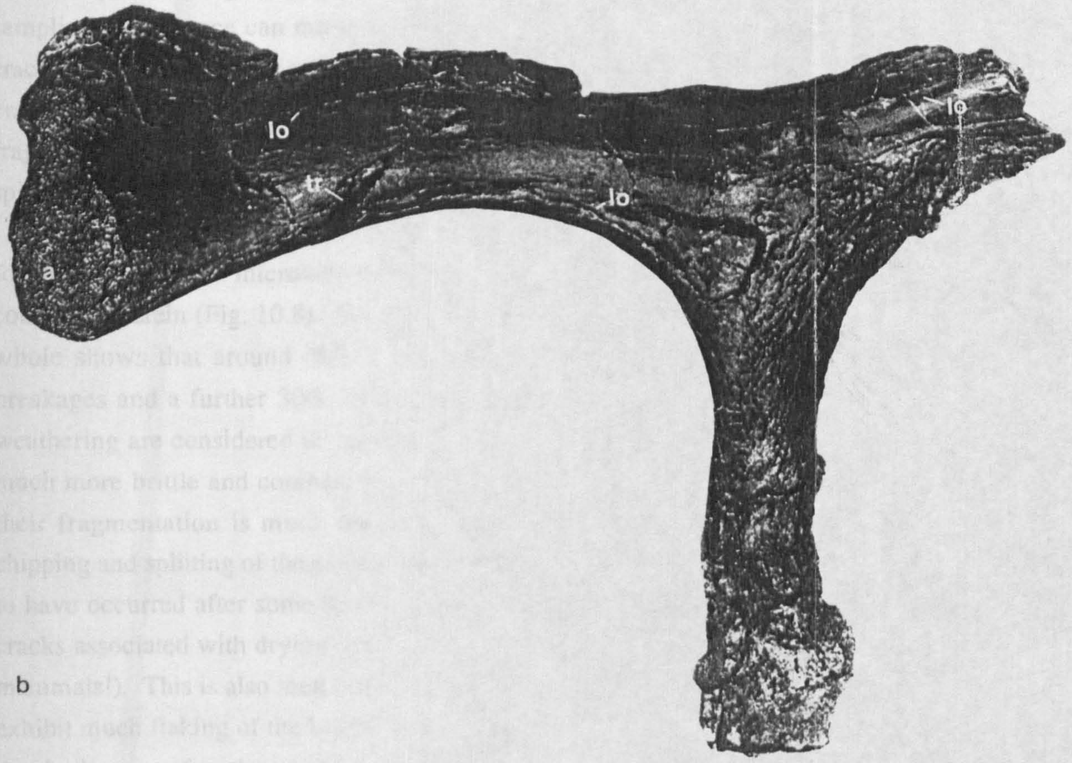
The main problem for a taphonomist is the practical identification of a fracture pattern for each individual broken bone, especially if that element shows several fractured edges. This can be done by comparing fragments of the same type of skeletal element in an assemblage. The second is to establish the agent which may have caused the breakage, although this can be alleviated by studying other surface features of the bone, such as looking for bite-marks or mineral growth (Johnson, 1985). This can also be countered by considering the fracture patterns for the assemblage as a whole, rather than for a single element, and comparing it with the patterns from accumulations were the agent of modification is known (Shipman, 1981).

10.2.1 Cetiosaur fracture patterns

Fracture patterns for the macrovertebrate remains seem to follow the general types described above and these are considered in greater detail in the relevant sections. All the bones examined showed a degree of fragmentation. Large scale breakages causing fragmentation of individual skeletal elements on the whole seem to have occurred when the bone was relatively fresh. Nearly 98% of the Hornsleasow macrovertebrate sample showed evidence of spiral fracturing, these breakages are located at the thin edges of bones, between processes and at the extremities of linear specimens (Fig. 10.7a-d). They are nearly always abraded, and show limonitic staining and/or mineral encrusting (Metcalf, 1993). They are most likely to have been caused by biological activity surrounding the bones in the hollow (sections 10.3 and 10.4). The bones are usually weathered and longitudinal cracks are commonly observed (section 10.5). Post-mineralisation fracturing is also common on the bones and around 85% of the studied bones showed transverse fracturing, although only 13 of these exhibited evidence of compaction shearing (Metcalf, 1993). This manifests itself as pervasive transverse cracking and small-scale faulting and squashing of the interior cortex of some bones (section 10.7.2). It also appears that the majority of the examined material (23 bones) was broken post-mineralisation by the explosive charges set during excavation (Metcalf, 1993).







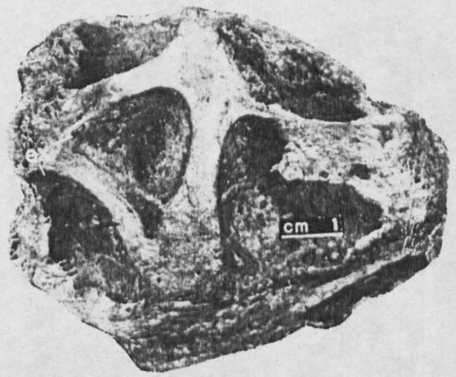


Figure 10.7. Fracture patterns exhibited by the cetiosaur remains: (a) and (b) right ilium fragment (G. 10001; lateral views), (a) shows the most weathered surface which lay 'face-up' in the hollow and has either been removed entirely ('wt' stage '2-3') or has deeply penetrating longitudinal ('lo') cracks running parallel to the grain of the bone. The bone also displays spiral fracturing ('sp'), and compaction squashing ('sq') and transverse fractures ('tr'). (b) is the more protected surface lying upon the karst, but shows some abraded (index '1') spiral breakages. (c) and (d) mid-posterior dorsal vertebra (G. 10000; anterior (c) and posterior (d) views), (c) centrum with unweathered, but broken surface, partial neural process exhibiting transverse cracking and limited spiral fractures. The latter have been abraded. (d) centrum shows more weathered surface (stage '2'), and substantial pre- and post-fossilisation breakage to the neural process. (e) a neural complex (G. 10131) exhibiting complete rounding (by abrasion and weathering) to a 'bone pebble'. 'ex' indicates where the surface of the bone (dark patches within recesses) has been exfoliated parallel to the inner fabric of the bone, giving the impression of an 'onion-skin' weathering pattern (stage '3').

10.2.2. Microvertebrate assemblage fracture patterns

Differentiating between spiral fracturing, and desiccation and post-mineralisation breakages is extremely difficult for small bone material. Most of the microvertebrate remains show some degree of chipping or breakage (Fig. 10.5a), but their small size makes elucidation of the cause extremely difficult (Figs. 10.8). On linear elements, compaction produces planar fractures, but on vascularised tissue, crushing of the interior causes irregular, splintering fractures. Also many of the small limb bones are hollow, therefore vertical static loading during compaction will succeed in breaking a mineralised bone along the grain of the fabric rather than perpendicular to it (pers. obs.). This will produce splinters similar to those produced by the dynamic action of crushing of fresh bone in a predators' jaws. If the bone is broken during extraction or preparation, the resulting 'transverse' breakage will be irregular and covered in 'bone dust' (pers. obs.), and hence, will resemble a pre-fossilisation spiral fracture, although the fracture surface tended to be of a lighter colour than the external bone surface and earlier fractures. Also if the bones are small, crushing will not only produce displacement along a crack, but may completely separate the two sides of the fracture. This will produce fragments in close proximity, but later biological, pedogenic, or even sampling disturbance can mask this effect (Maas, 1984). Lastly, even longitudinal cracks associated with desiccation might cause confusion for really tiny elements. If the cracks are deeply penetrating to cause separation along them, a linear element might fragment into splinters, roughly equal in shape and proportion to those produced by spiral fracturing.

Nevertheless, I have attempted to differentiate between different fracture patterns for the Hornsleasow microvertebrate assemblage (Fig. 10.5l) and different elements contained therein (Fig. 10.8). The general pattern (Fig. 10.5l) for the assemblage as a whole shows that around 40% of the vertebrate material exhibit pre-fossilisation breakages and a further 30% show transverse fracturing (longitudinal cracking and weathering are considered in section 10.5). Enamel covered teeth (Fig. 10.8a) are a much more brittle and compact skeletal material than bone. Therefore, diagnosis of their fragmentation is much more difficult to determine, for instance, conchoidal chipping and splitting of the enamel could happen at any time, although it is most likely to have occurred after some desiccation (caution is suggested though as longitudinal cracks associated with drying out of enamel has been observed by the author in living mammals!). This is also seen in the examination of fish scales (Fig. 10.8g), which also exhibit much flaking of the layered enameloid structure during desiccation. It appears that in the case of teeth examined much of the breakages which has occurred at the root or base is caused by spiral fracturing (e.g. Fig. 10.5d), possibly associated with the

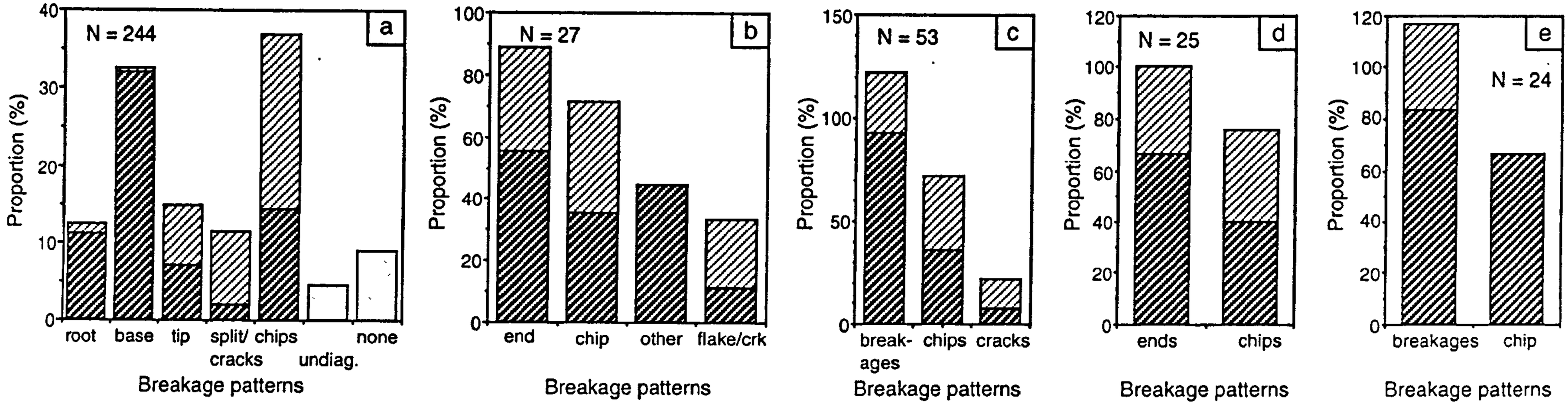


Figure 10.8. Breakage patterns on vertebrate material (a) teeth; (b) jaws; (c) skulls; (d) ribs; (e) vertebrae; (f) dermal armour (including carapace); (g) fish scales; (h) limb bones; & (i) dermal bone, including girdle elements, showing percentage with spiral and transverse fractures. Number (N) is indicated.

transverse spiral undiagnosed

loss from the jaw. Spiral breakage of the thinned edges of scales seems to be related to digestion damage (section 10.3).

Linear, compact elements with little internal cancellous tissue, such as jaws (Fig. 10.8b), ribs (Fig. 10.8d) and limb bones (Fig. 10.8h), exhibit similar fracture patterns and diagnosis of their types was much more successful. Most of the breakages occurred through the main axis of the shaft (e.g. Fig. 10.5f,g,i), and of these two-thirds to three-quarters were spiral in nature (Fig. 10.8b,d,h). However, if the fracture had occurred through the articular end (e.g. Fig. 10.5c), the ragged nature of resulting breakage could be more to do with the microstructure of the cancellous tissue contained therein, rather than a predominance of fresh bones being broken at their articular ends. Cracking and flaking seems to be associated with weathering of dried bones. The subequal proportions of spiral and transverse chipping of linear bones, may reflect the difficulties in elucidating their cause.

Bones containing much cancellous tissue, such as skull elements (Fig. 10.8c), vertebrae (Fig. 10.8e), dermal armour and carapace material (Fig. 10.8f) and dermal bone and girdle bone (Fig. 10.8i), also exhibit similar fragmentation patterns. Generally, fractures tend to be ragged or irregular, reflecting their complex internal microstructure (e.g. Fig. 10.5e,k). Spiral fracturing seems to be most prevalent at the edges of dermal bones or upon processes, where chipping is also observed. Deeply penetrating cracks are usually roughly smooth sided and often associated with mineralisation, these seem to have formed post-fossilisation.

10.3. Primary taphonomic sorting - biological modification of bones

Primary sorting is associated with the death of the animal or faunal community. The death may occur as a result of an accident, it might be deliberate (i.e. predation), or it may occur due to 'natural' causes (e.g. old age or sickness). The taphonomic processes and agents which cause primary taphonomic sorting include disarticulation and breakage resulting from the action of predators or scavengers (e.g. Mellet, 1974; Dodson & Wexlar, 1979; Korth, 1979; Pratt, 1979, 1989; Andrews & Nesbit-Evans, 1983) and the effects of stomach acids on tooth enamel or bone (Mayhew, 1977; Fisher, 1981; Andrews & Nesbit-Evans, 1983). These usually leave reasonably distinguishable features upon the remains of the animal.

Bone breakage and fragmentation of skeletons can be caused by carnivore activity. Studies of modern assemblages indicate that fresh bone material breaks with a spiral fracture. Spiral fracturing always occurs prior to any alteration of the bone material, and is generally caused by chewing and splintering by predator action (Hill, 1980; Fiorillo, 1988a). An abundance of spiral fractured elements and similar-sized fragments in an accumulation may indicate predator activity. However, spiral fracturing can also be caused by non-biological or secondary biological processes which occur

prior to desiccation or fossilisation of the bone, such as breakage by transport, roll or trampling (Fiorillo, 1988a). Spiral fractures may be abraded and worn by erosion in transported bones (section 10.6).

As all of the carnivores in the Hornsleasow assemblage bore teeth, an investigation into the amount and degree of carnivore utilisation of the remains of prey or scavenged items was undertaken, in assessing the presence or absence of tooth or gnaw marks on the bones. Tooth marks and scrapes are common in modern mammalian carnivore-modified vertebrate assemblages. They usually identified as fairly deep grooves, U-shaped in cross-section often running perpendicular to the axis of the bone or as deeply penetrating puncture marks, with associated radial cracking around the hole (Hill, 1979b; Haynes, 1980a,b, 1982; Binford, 1981; Fiorillo, 1988a). These marks may be isolated or may occur in multiple sets. Other associated marks include gnaw-marks, these give the bone a characteristically 'chewed' or ragged look, in many mammalian assemblages bones may show 'scalloped' edges which indicate gnawing by carnivores (Fiorillo, 1988a). Gnaw marks are also known to be made by herbivorous mammals, some such as rodents, which utilise bones to wear down their continuously growing incisors, and produce characteristic deep paired grooves upon bones (Brain, 1981; Fiorillo, 1988a; 1991). Chewed bones are also often crushed (showing spiral fracture patterns) and can become polished, especially by the abrasive action of licking and sucking by mammalian carnivores (Haynes, 1980a,b). Attention should be paid to the proximal ends of bones, which are often gnawed or crushed by mammalian predators in order to utilise the nutrients available from the bone apatite or the pulpy interior of the bone. Larger carnivores are known to consume portions of their prey bones or to swallow their prey whole in order to utilise this nutritional source (Fiorillo, 1991). Carnivores may prefer particular skeletal elements or parts of them, for example, modern mammalian predators seek out the pulpy tissue enclosed in long bone epiphyses (Shipman, 1981). Therefore preferential destruction of one portion of a bone might also be characteristic of damage caused by a particular carnivore.

Whereas Haynes (1980a,b, 1982), Hill (1980) and Fiorillo (1988a, 1991) studied breakage and bone surface modification on the remains of large vertebrates, breakage patterns and other features of carnivore utilisation of modern and fossilised microvertebrate remains have also been studied by other authors. Actual ingestion and digestion of microvertebrate remains leaves distinctive traces, such as stripping of the enamel upon teeth, and dissolution and thinning of bones (e.g. Mayhew, 1977; Fisher, 1981). These can readily be discovered through SEM investigation of bone material. Many studies of microvertebrate bone accumulations made by modern carnivores have been made, and include investigations of coprocoenosis made mammalian predators (Mellet, 1974; Korth, 1979; Andrews & Nesbit-Evans, 1983), birds of prey (Mayhew, 1977; Dodson & Wexlar, 1979; Korth, 1979; Worthy & Holdaway, 1994), crocodiles

(Fisher, 1981) and Komodo monitor lizards (Auffenberg, 1981). Certain features of these accumulations can serve to identify the type of predator involved in forming the accumulation. For instance, crocodiles and the Komodo monitor tend to demineralise the bones of their prey completely during digestion, especially those of small prey items (Fisher, 1981; Auffenberg, 1981). However, conversely, other authors (e.g. Weigelt, 1929) noticed that crocodiles often regurgitate hard prey items, including bones and teeth and if feeding is high, then the faecal material of Komodo monitors is known to contain undigested bone material (Auffenberg, 1981). Therefore, in many cases it is still difficult to classify modern and Tertiary microvertebrate accumulations in terms of predator identity (Maas, 1984).

Indirect evidence for primary taphonomic modification of bone material in and around the pond comes from the abundance of fossilised carnivoran faecal pellets. The association of a carnivorous reptile's shed or broken teeth with the remains of a presumed prey item, can also suggest utilisation of a carcass (Buffetaut & Suteethorn, 1989) and the presence of breakages or wear upon carnivore teeth may suggest contact between tooth and hard-prey items (Fiorillo, 1991).

10.3.1. Evidence for predation and scavenging in the macrovertebrate assemblage

Biological processes such as disarticulation, scattering and breakage of the cetiosaur skeleton are largely a function of predator and scavenger activity (cf. Hill, 1979b; Haynes, 1980, 1982). Indirect evidence for the utilisation of the carcass by carnivores is indicated by four large serrated teeth (over 50mm length) of the theropod *Megalosaurus*, which were recovered in close proximity to the bones (see Fig. 10.10c; Metcalf, 1993). These teeth were excavated from the base of the lens, and are all broken crowns, or with partial, broken roots, rather than shed teeth. Teeth of other small theropods (5-10mm in total length), and crocodiles were also found within the basal clay unit during sieving, but as this process mixes recovered skeletal material from all depths within the clay context being sieved, and it is difficult to decide whether these represent teeth from scavengers upon the cetiosaur, or from higher up in the Grey Clay paleosol.

The condition of the bone was also studied for evidence of biological activity. Around 30% of the cetiosaur remains showed evidence of carnivore activity, including four bones with shallow paired grooves, and one exhibiting paired puncture marks (with associated radial cracking). These traces are interpreted as tooth scrapes and bite marks. They are roughly in proportion to the size of the teeth of small reptilian carnivores, such as ceratosaur theropods or goniopholid crocodiles.

Although many of the bones are broken and fragmented (Fig. 10.7a-b), it is difficult to elucidate whether some of the pre-fossilisation fracturing was caused by

splintering in the jaws of carnivores (Haynes, 1980, 1982), or by trampling and/or roll within the hollow.

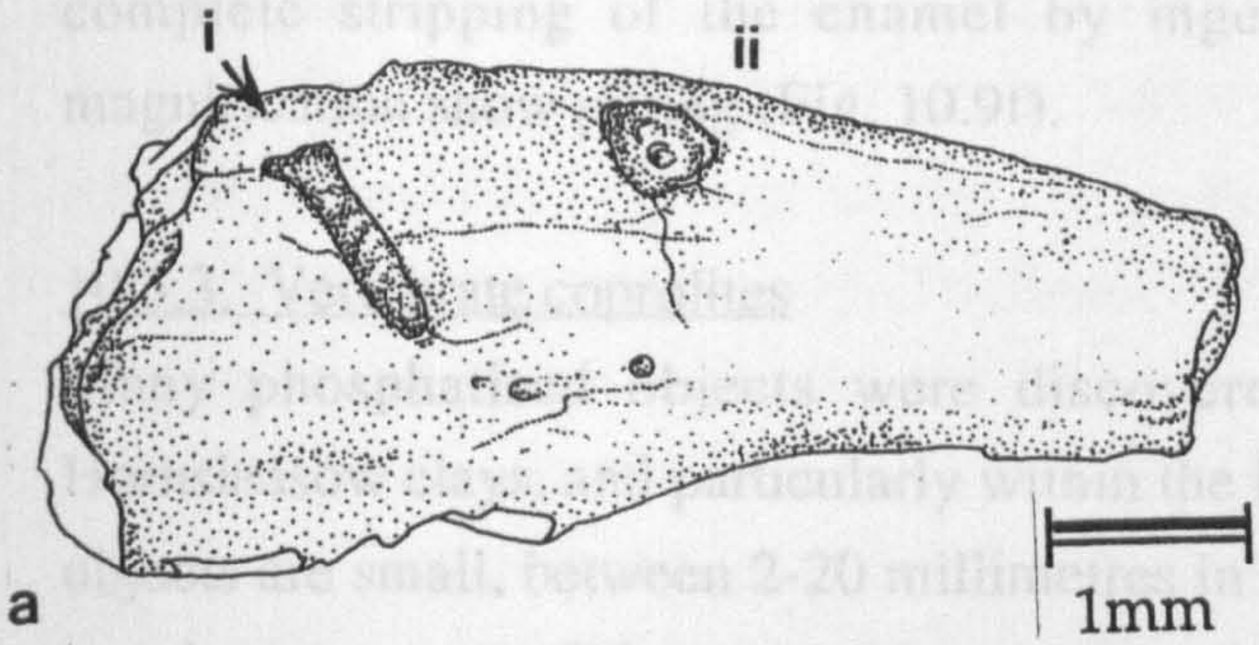
10.3.2. Evidence for predation and scavenging in the microvertebrate accumulation

The main cause of sorting within a microvertebrate community, would be by the effects of predation. Although the Hornsleasow remains are fragmentary, making biological reconstruction difficult, examination of the more complete elements in the fauna and comparison with those from other Middle-Upper Jurassic sites would suggest that most of the animals were under two metres in length. They include the amphibians (Evans *et al.*, 1988, 1990), the lepidosauromorphs (Evans, 1991), the champsosaur *Cteniogenys* (Evans, 1989, 1990), the crocodilians (Evans & Milner, 1992), the ornithopods (Thulborn, 1973), and 'ceratosaur' theropod dinosaurs (pers. obs.), the mammal-like reptile *Stereognathus* (Waldman & Savage, 1972) and the mammals (Freeman, 1976, Savage, 1984). Therefore, if the accumulation represents a coprocoenosis the predator/prey size is roughly constrained. Elements from larger beasts are rare (excluding the basal *Cetiosaurus* remains, and the teeth of associated scavengers) and therefore are likely to be exotic elements introduced into the assemblage by other means of accumulation.

It is possible that the severe fragmentation of elements at Hornsleasow is a result of predator activity, although in the absence of direct evidence (bite marks, stripping due to stomach acids) and modern analogues, it is difficult to prove for all but a few remains. Only 0.5% of all bones studied showed evidence of bite marks, in the form of single or occasionally paired grooves. One bone, the tiny fragment of an archosaurian humerus (G. 55963) showed paired bite-marks consisting of a U-shaped groove with small perpendicular ridges produced by the "chattering" of the tooth across the surface of the bone and associated puncture mark (Fig. 10.9a,b). The ends of the humerus have been spirally fractured and it is likely that this happened at the same time as the biting. The tooth marks are fairly small (less than 0.5mm in diameter) and were probably caused by a microvertebrate carnivore, such as a goniopholid crocodile. No scalloped edges attributable to chewing and no gnaw marks were seen on any of the bones examined.

There is evidence for ingestion of small vertebrate remains by predators. Acid digestion damage is seen on around 10% of the sample studied. This manifests itself as pitting, stripping and pock-marking of the surface of bones and enamel covering of teeth. Thinning of the extremities or broken edges of skeletal remains is also observed. Many of the *Lepidotes* scales are extremely worn, showing thinning at their edges and a general pitting and dissolution of the enameloid covering (Fig. 10.9c). A few of these scales were found adhering to the surface of, or in close proximity to, small pelletal coprolites - suggesting that the fish may have been the diet of a small predator in the

Figure 10.9. Primary bone modification. Bite marks: (a) and (b) Drawing and SEM image of an archosaurian humerus fragment (G. 55963, lateral view), showing paired bitemarks: (i) stepped groove with cracks, and (ii) puncture with radial cracks. Digestion damage: (c) G. 60029, fish scale exhibiting pitting and stripping of enameloid layers (field of view = 1.3mm); (d) G. 51220, surface of crocodile tooth, showing stripping of enamel ('el') and pitting of the underlying dentine ('dn') (field of view = 0.4mm); (e) G. 60904, crocodile tooth fragment showing stripped enamel (tooth length = 1.6mm); (f) G. 50613, remnant crocodile tooth showing complete stripping of enamel and erosion of underlying dentine (tooth length = 3.4mm). Vertebrate coprolites: (g) G. 84001, bullet shaped pellet (length = 10.5mm); (h) G. 73286, sausage shaped pellet (length = 14.4mm); (i) and (j) G. 73300, buckled sausage shaped pellet (length = 35mm), (i) photograph and drawing constricted surface and mucro ('mu'); (j) photograph of surface showing coprophage scratches ('s'); (k) G. 93291, cross-section through pelletal coprolite showing radial fractures in the centre (width = 9.8mm).



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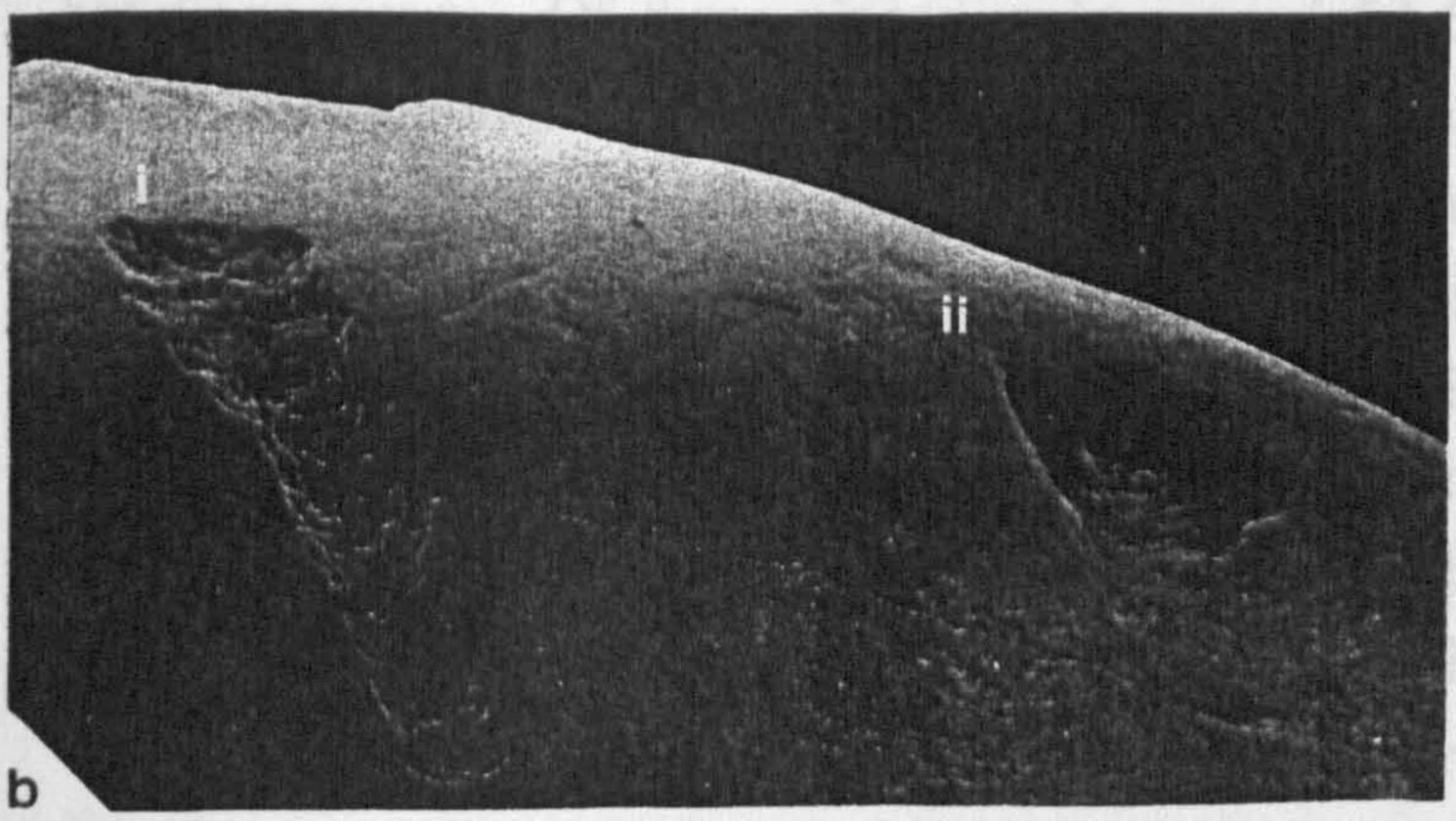

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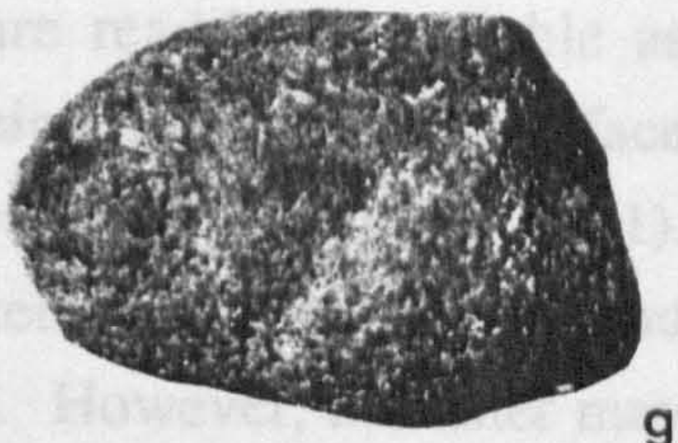

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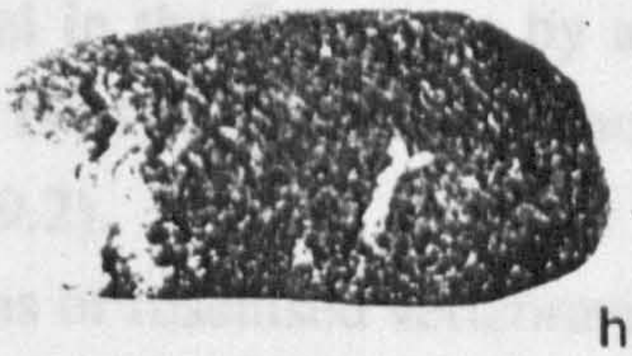

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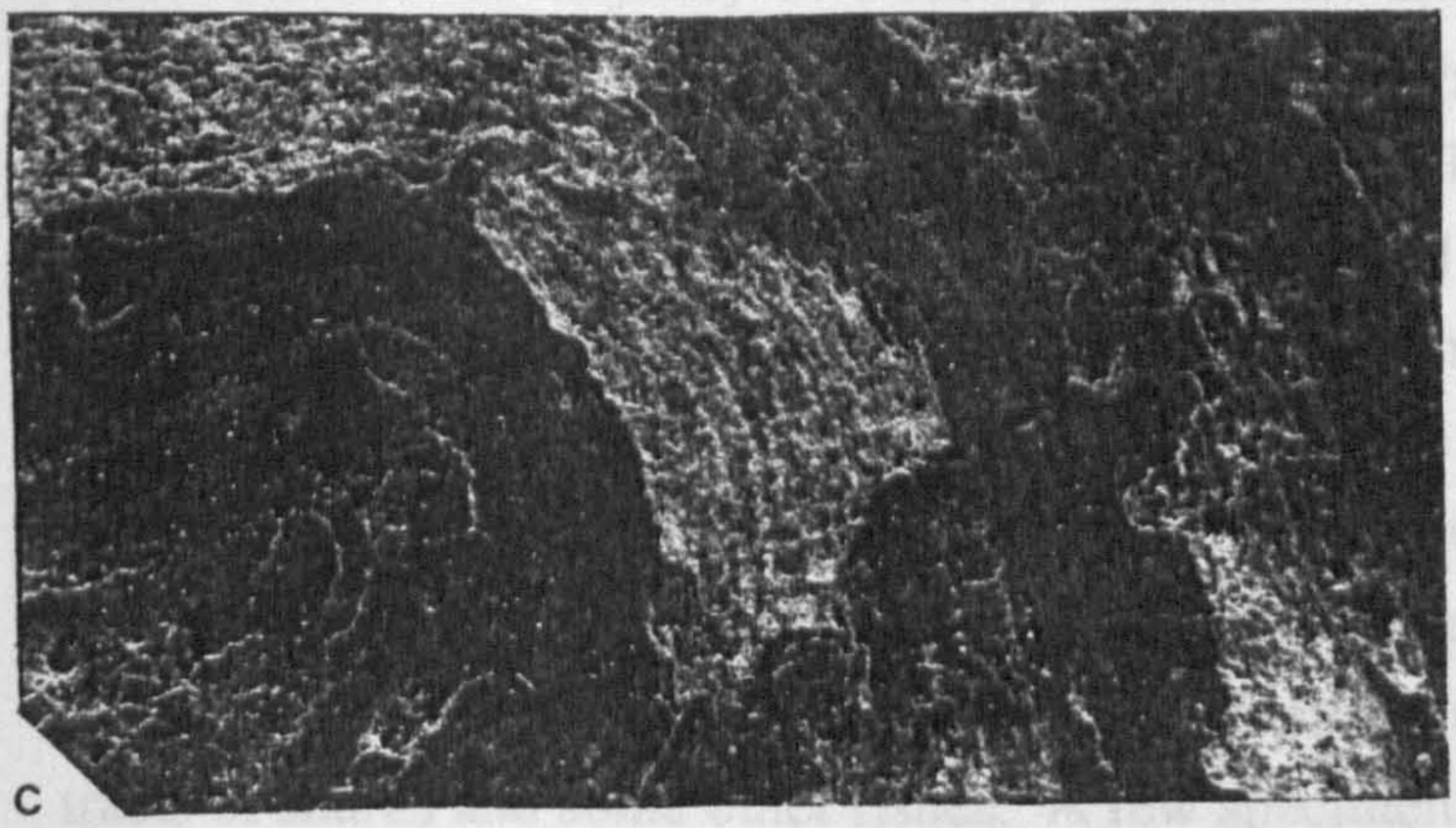

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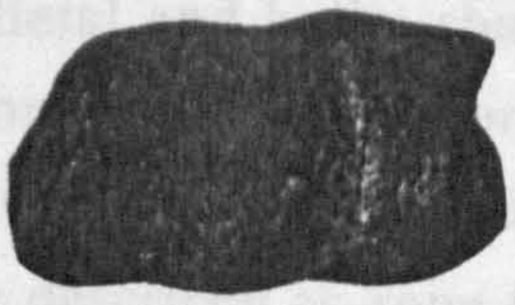

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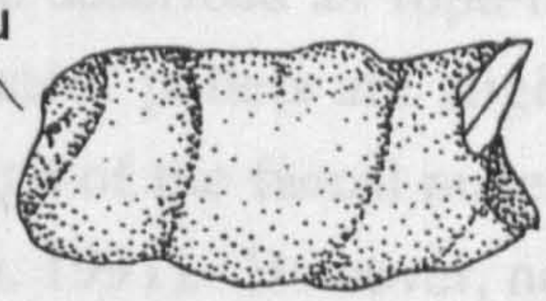

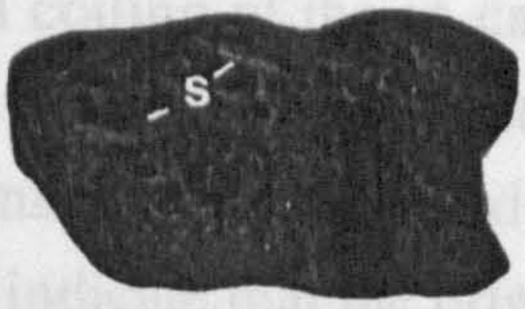

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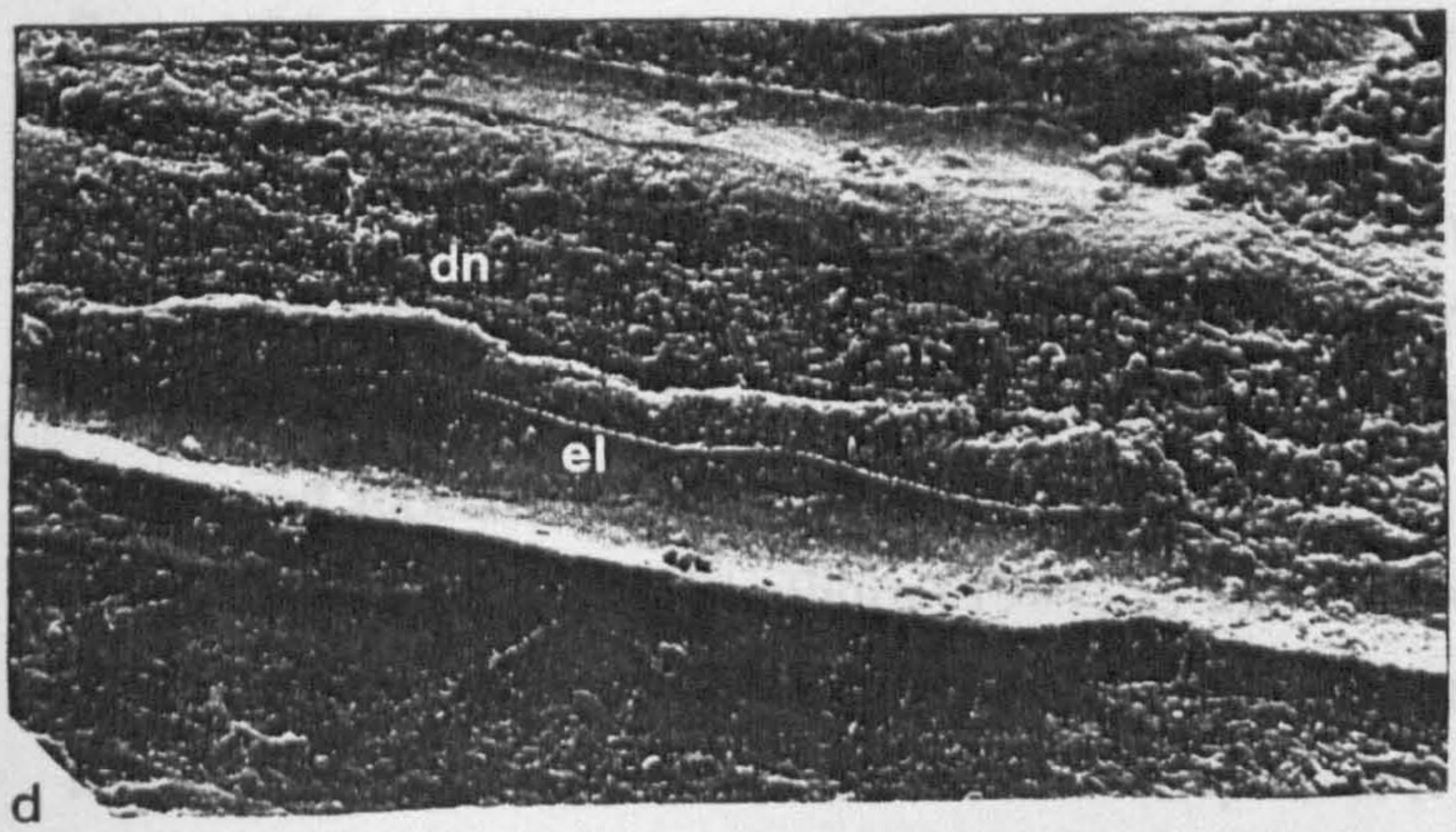

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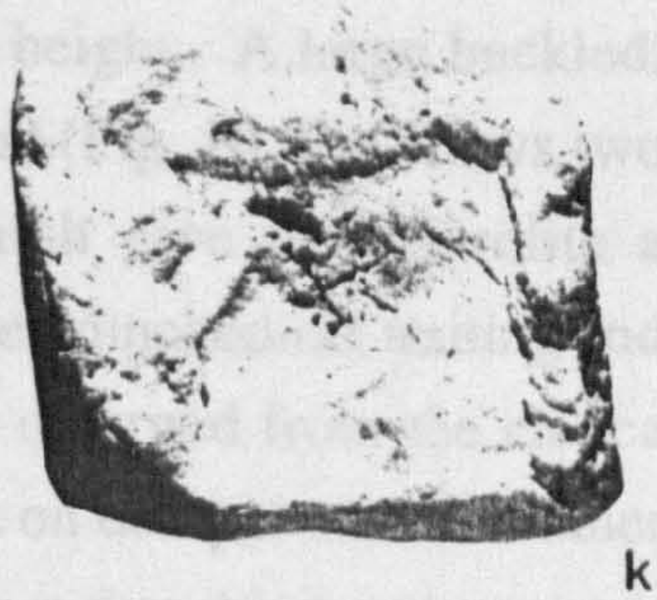

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pond (?crocodile: Metcalf, 1993). Some of the archosaurian teeth also show partial or complete stripping of the enamel by ingestion (Fig. 10.9d,e), which at high magnification show pitting (Fig. 10.9f).

10.3.3. Vertebrate coprolites

Many phosphatised objects were discovered within the sieved residue of the Hornsleasow clays, and particularly within the lower Grey Clay. The majority of these objects are small, between 2-20 millimetres in maximum breadth, and regular shaped, but fragmentary. Of these pieces, only around 50-60 are readily identifiable as fossilised vertebrate faecal matter or 'coprolites', from their distinct shapes, surface textures, composition and the presence of inclusions or adhesions (cf. Thulborn, 1991). The rest are probably a mixture of non diagnostic or broken coprolitic material and inorganic pedogenic phosphate nodules (see section 5.9.2). However, the latter may have nucleated and grown around phosphatic faecal material in the first place by a process known as prefossilisation (Antia, 1979, Thulborn, 1991) and so cannot be readily distinguished in some cases (see below, and section 5.9.2).

Of those objects which are considered to be the remains of fossilised vertebrate faecal matter, the majority of these specimens are small pelletal and bullet-shaped masses (e.g: specimens G. 84001, 84002: Fig. 10.9g). Some are quite strongly elliptical and can be described as 'sausage-shaped' (Thulborn, 1991; e.g. G. 73286: Fig. 10.9h), whilst others are much more elongated and can be described as 'rope-like' (Thulborn, 1991; e.g. G. 7.3287). The majority of these faecal pellets are slightly curved or coiled, this is thought to be a function of the passage of the faecal material through the intestine and extrusion from the cloaca (Thulborn, 1991). However, none of the Hornsleasow specimens show the intricate folding and coiling of the so-called 'spiral coprolites' (Williams, 1972), which are attributed to the fossilised intestinal tracts of sharks and some other fishes. A few specimens are much more squashed and flattened in appearence (e.g. specimen G. 73288), which may indicate that the original droppings were much softer and unconsolidated than those which exhibit a definite regular shape or that they were dropped from a far greater height. A large buckled, sausage-shaped specimen (G. 73300; diameter 12 millimetres) (Fig. 10.9i,j) shows two well-defined constrictions in width along its length, which give the coprolite a segmented appearence. This specimen also exhibits the pointed pinched-off trailing end or 'mucro' described by Thulborn (1991) as that end which emerged from the cloaca last. The other end of the specimen is irregularly broken, but on comparison with other fragmentary specimens and Thulborn's descriptions, it appears that this breakage may have happened during defecation (rather than post-mineralisation). Therefore, it would appear that this coprolite represents the fossilised remains of one of a series of stubby, block-like faecal pellets, much like horse manure (cf. Thulborn, 1991, fig. 7h).

The Hornsleasow coprolites show a variety of surface features. All of the definite faecal pellets, exhibit a thin darkened 'halo' on the outer and unbroken surface. In some specimens (e.g. G. 73300, G.84001, G. 73287) the surface also exhibits a smoothed and almost glossy lustre. The darkened halo and smoothed appearence can be attributed to several factors (Thulborn, 1991): firstly, it may be the remains of an original mucous coating on the surface of the dropping (Matley, 1941); secondly, it could be chemically altered faecal material or sediment which surrounded the dropping (Thulborn, 1991); and finally, it could be attributed to a secondary mineral coating, such as iron oxide or limonite (Thulborn, 1991). It appears that there might be a light dusting of limonite upon many of these specimens, but the halo in most cases appears to be texturally similar to the rest of the coprolite and therefore can be considered to be a primary coating. Interestingly, both ends of the buckled, sausage-like pellet (specimen G. 73300) exhibit the darkened halo indicating that the irregular shape is in fact original rather than post-depositional.

None of the coprolites show the surface striations or grooves described by Thulborn (1991), however, many do show fine-scale surface pitting, which are the product of gas bubbles generated by decomposition (Thulborn, 1991). Gas vesicles are also seen in freshly cut or broken cross-sectional fragments (pers. obs.). One specimen, G. 73291, shows small-scale radial fissures within the broken interior (Fig. 10.9k). Similar cracks have been described and figured in Thulborn's paper and are described by him as shrinkage or desiccation cracks. One surface of specimen G. 73300 shows short (1-2mm), fine, and straight-sided marks (Fig. 10.9j). These are similar to those shown by Thulborn (1991, fig. 7d) and which are derived from the action of coprophages (such as dung-beetles), micro-organisms or fungi. Those upon the Hornsleasow specimen most closely resemble the scratches made by coprophages in both morphology and size.

Some of the smaller coprolites appear to contain microscopic inclusions of undigested organic material within their interior. Most of this material appears to be tiny fragments of bone and fish scale, and no plant remnants were seen within the Hornsleasow coprolites. Other coprolite specimens were found closely associated with pieces of bone and scale, and in a few cases fish scales were found adhering to the surface of the pellet. The detritus is thought to have become embedded within the surface of the moist faecal pellets, whilst they were rolling about within the sediments (Thulborn, 1991). The groundmass of the coprolites are texturally amorphous.

The chemical composition of the Hornsleasow coprolites as calcium hydroxyapatite and the inclusions which they contain, suggest that they were extruded by carnivorous or piscivorous animals. The faecal pellets of carnivorous animals are much more likely to be fossilised than those of herbivores, as they are rich in calcium phosphate and therefore need little mineralisation to be preserved (Retallack, 1984).

Those of herbivores would contain digested plant matter, which would tend to be preserved as coalified material, if at all.

The survival of the coprolites within the Hornsleasow paleosols can be explained in terms of the general taphonomic environment which prevailed in the two main clay layers. The excellent preservation of phosphatic material (i.e. coprolites and fossilised bone material) within the lower Grey Clay is problematic, considering that the depositional environment was within a rather de-oxygenated pool. Retallack (1984) found that phosphatic coprolites were absent from paleosols which were presumed to have been acidic and the preservation of unoxidised wood material within the clays, would suggest that the Grey Clay conditions were locally acidic. In fact, some bone material exhibits pitting and pock-marking from humic acid erosion (section 10.4.2), but this is not seen upon the surface of the phosphatised coprolites from this layer. However, the current-free, de-oxygenated waters of modern swampy areas are known to inhibit biological decay and physical destruction of faecal matter (Retallack, 1984), and a similar situation may have been true for the Hornsleasow coprolites.

Phosphatic coprolites are also well-preserved within the aerated, well-drained soils of the upper Green Clay at Hornsleasow. These specimens are much more likely to show desiccation cracks (e.g. G. 73291) and it has been suggested that the soils dryness might have been an important factor in retarding microbial and chemical degradation of the faecal pellets (Retallack, 1984). Contrary to popular belief, intact faecal matter cannot be transported very far from source (Korth, 1979) and therefore it is likely that the coprolites found within the Hornsleasow soils are derived from animals defecating within or in the region of the doline (Metcalf, 1993).

The provenance of individual Hornsleasow coprolites is difficult to elucidate. For instance, unrelated animals can produce droppings which are virtually identical in form, whilst related animals, and even the same individual, can produce quite distinct droppings depending on their physiological or psychological state (Thulborn, 1991). Therefore, although there are several distinct coprolite morphologies represented at Hornsleasow, it is almost impossible to say 'who dung what'! However, there are a few general statements that can be made about the animals responsible. Firstly, the smaller pellets are probably derived from small animals, usually the diameter of the individual coprolite is roughly equal in size to that of the animals' intestinal tracts and cloaca. Although, this is not always the case, for example, both goats and rabbits produce similar sized pellets (Thulborn, 1991). The largest identified coprolite in the assemblage is the buckled sausage-shaped specimen (G. 73300) is only 12 millimetres in diameter, although it appears that this is only part of a string of droppings. As some of the coprolites contain fish material, it is highly probable that these were produced by aquatic-adapted animals, such as the many small predatory reptiles (Fisher, 1981) for example the goniopholid crocodiles, the champsosaur *Cteniogenys*, or the

pleurnosternid turtles. They could also be derived from the spear-toothed, piscivorous rhamphorhynchoid pterosaurs or even small theropod dinosaurs. Coprolites of terrestrial carnivores (maniraptoran theropods) may have also been introduced into the hollow directly (i.e. when the animal came to drink or hunt at the water's edge) or indirectly by flooding events. Much of the interior of the coprolites represents ground-up material with very little recognisable inclusions, this may suggest that the Hornsleasow predators had an extremely acidic gut (cf. crocodile, Fisher, 1981) or that they contained stomach stones or gastroliths. However, no gastroliths have been recognised within the assemblage.

10.3.4. Discussion

Although it is common in modern and Tertiary bone accumulations to find evidence of utilisation by mammalian carnivores, it is rare within predominantly reptilian assemblages (Fiorillo, 1991). Predator utilisation of bone material in Mesozoic reptilian assemblages has been documented by tooth mark traces upon bones by Clark & Etches (1991: marine reptiles), Dodson (1971: dinosaurs) and Fiorillo (1991: dinosaurs), and from the indirect evidence of the association of carnivoran teeth with potential prey items by Buffetaut & Suteethorn (1989: dinosaurs). In all these assemblages utilised prey bones are observed in much lower frequencies (comparable to those seen at Hornsleasow) than in mammalian assemblages. Fiorillo (1991) suggests that this is because mammalian predators are much more suited to crushing bones than reptiles. However, not all mammalian assemblages contain high frequencies of bitten bone - for instance in times of hardship a prey carcass is much more heavily utilised by mammalian carnivores than when hunting is good (e.g. Haynes, 1980a,b) and the amount of carcass utilisation also depends upon the length of exposure before burial (Shipman, 1981). However, the time-averaged assemblage at Hornsleasow should average out this type of seasonal variation.

The Hornsleasow assemblage is largely made up of an extinct reptilian community, hence any predator/prey activity can only be surmised. For instance, although predatory reptiles of the Mesozoic bore teeth, they seem to have not routinely crushed or chewed their prey item's bone in a manner similar to that of many mammalian carnivores. However, they may have utilised the nutritional value of bone material by swallowing their prey whole and digesting the hard-parts along with the flesh (cf. crocodiles, the Komodo monitor or raptorial birds) - prey bones being passively consumed with their prey carcass, rather than desired as a means of obtaining nutrients (Auffenberg, 1981; Fisher, 1981). If modern reptiles swallow larger bones then these are not crushed or gnawed, and therefore tooth marks only occur through glances with the teeth during ingestion.

It is probable that the crocodiles, small lepidosauromorph reptiles and the turtles represented in the fauna, behaved in a similar manner to their modern counterparts, but in the case of the extinct beasts, it is much more difficult to provide the answers. For instance, did small theropod dinosaurs have dining habits similar to crocodiles and the Komodo monitor, or were they more akin to raptorial birds? Fiorillo (1991) speculated that theropod dinosaurs may have not been able to chew bones, and that carcass wastage may have been quite high for these carnivores. He even suggested that they may have been primarily scavengers, utilising rancid meat on decaying carcasses. He suggested that this could be because the larger prey items, such as other dinosaur cadavers, may have possessed bones too robust for chewing or that the morphology of the jaws and teeth of carnivorous reptiles was not sufficiently robust to break bones. Theropods like other reptiles routinely shed their teeth, and therefore contact between tooth and prey bone may have dislodged teeth from the jaw. Fiorillo (1991) also stated that most of the theropod teeth he observed bore no wear-facets or tip-breakages and concluded that bone crushing was not routine amongst predatory dinosaurs. He suggested that theropods probably had similar digestive tracks to Komodo monitors or crocodiles and therefore may have obtained nutrients from swallowing small bones whole. However, about a third of the *Megalosaurus* (Fig. 10.10c) and small theropod teeth examined at Hornsleasow do show wear facets (Fig. 10.10d), as do over half of all the crocodile teeth (Fig. 10.10a,f-h) and pterosaur teeth (Fig. 10.10b) examined. Therefore, the Hornsleasow predatory reptiles were occasionally consuming food items which caused breakage and wear of their teeth, without significant discomfort or loss, even though some teeth were so broken that they resembled worn nubs (Fig. 10.10h). This evidence suggests that theropods and other Hornsleasow reptiles may have utilised prey bones.

All the signs of bone utilisation can be attributed to carnivore activity as although primitive mammals are found in the Hornsleasow assemblage, no paired grooves similar to rodent gnaw-marks were found and none have so far been reported from similar Mesozoic accumulations (Fiorillo, 1991). All the evidence outlined above indicates that the pond would have attracted many animals as a watering hole, and there would be ample opportunity for predation and scavenging. The pond may have formed a residence for crocodiles and other small carnivores, as modern crocodilians maintain small ponds and watering holes, in which they can survive periods of drought, and use as effective prey traps (N. Rutger, pers. comm., 1992). Fiorillo & Hutchison (1992) have recently reported such fossil crocodile ponds in the Tullock Formation (Palaeocene) of Montana.

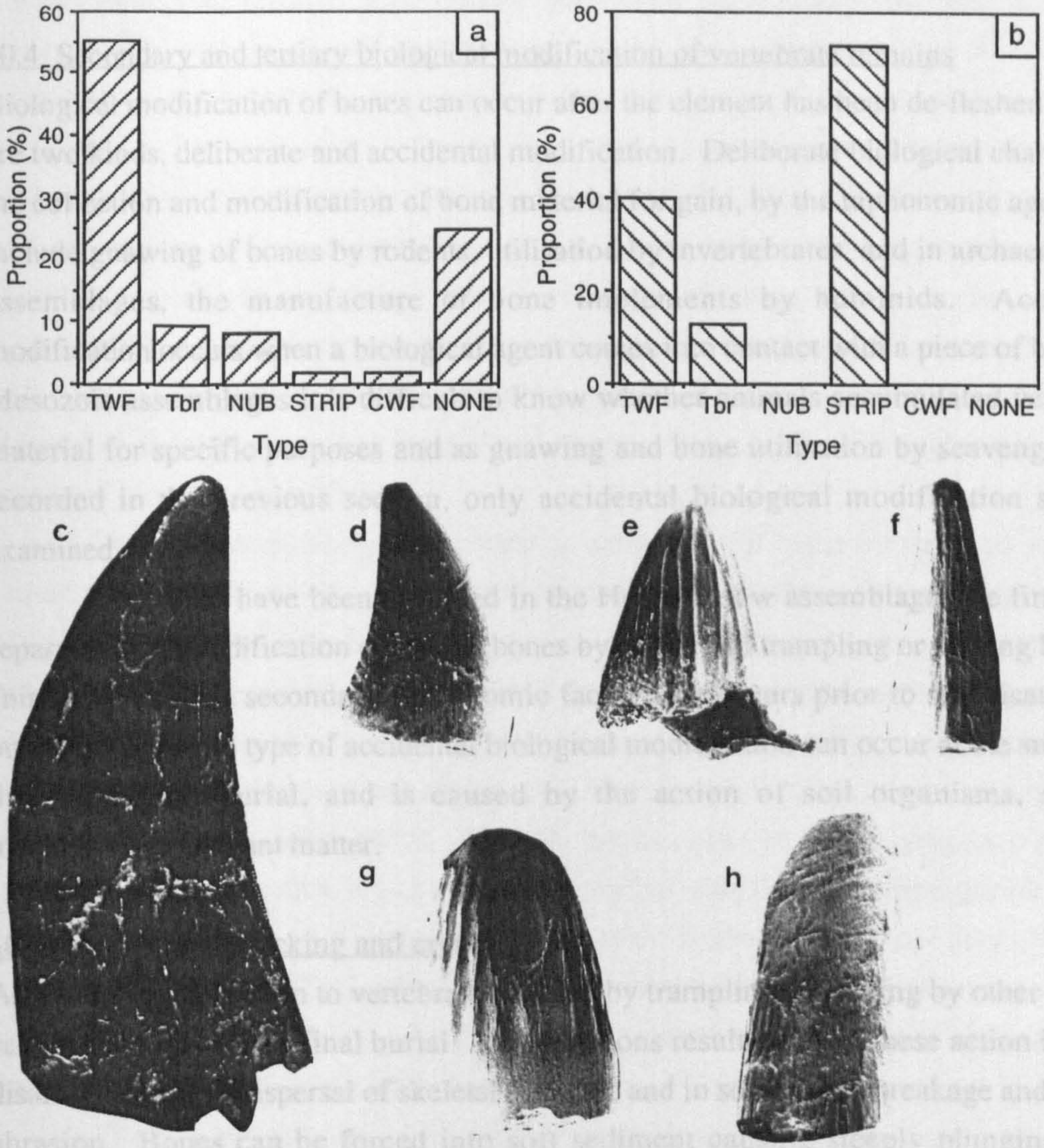


Figure 10.10. In life wear on teeth in the microvertebrate assemblage. Proportions and types of wear or breakage on (a) crocodile and (b) pterosaur teeth. Abbreviations: 'TWF' tip wear facet; 'Tbr' tip breakage; 'STRIP' stripped wear facet; 'CWF' carinae wear facet. Details of wear on teeth: (c) GMHM1, *Megalosaurus* shed crown with large tip wear facet (crown height = 68.3mm); (d) G. 60301, small theropod shed crown with tiny tip wear facet (crown height = 1.5mm); (e) G. 61403, crocodile broken crown with no wear facets (crown height = 1.4mm); (f) G. 51220, shed crocodile tooth with flat-topped tip wear facet (crown height = 3.7mm); (g) G. 54248, shed crocodile tooth with large tip and facial wear facet (crown height = 3.1mm); (h) G. 61616, worn nub of a crocodile tooth (crown height = 1.7mm).

10.4. Secondary and tertiary biological modification of vertebrate remains

Biological modification of bones can occur after the element has been de-fleshed. There are two kinds, deliberate and accidental modification. Deliberate biological changes are the collection and modification of bone material for gain, by the taphonomic agent, and include gnawing of bones by rodents, utilisation by invertebrates, and in archaeological assemblages, the manufacture of bone implements by hominids. Accidental modification occurs when a biological agent comes into contact with a piece of bone. In Mesozoic assemblages it is difficult to know whether animals accumulated vertebrate material for specific purposes and as gnawing and bone utilisation by scavengers was recorded in the previous section, only accidental biological modification shall be examined.

Two types have been recorded in the Hornsleasow assemblage, the first is the separation and modification of animal bones by accidental trampling or kicking by other animals. This is a secondary taphonomic factor as it occurs prior to stabilisation and burial. The second type of accidental biological modification can occur at the surface or during shallow burial, and is caused by the action of soil organisms, such as invertebrates and plant matter.

10.4.1. Trampling, kicking and crushing

Accidental modification to vertebrate remains by trampling or kicking by other animals can occur prior to their final burial. Modifications resulting from these action includes disarticulation and dispersal of skeletal remains, and in some cases breakage and surface abrasion. Bones can be forced into soft sediment causing steeply plunging burial attitudes not normally associated with normal methods of deposition (Hill & Walker, 1972). Trampling can also cause spiral fracturing damage and chipping of fresh bone material and small or weathered bones may be broken completely (Myers *et al.*, 1980), although Andrews & Cook (1985) suggest that complete breakage is not at all common in macrovertebrate remains. Localised chipping and spiral fracturing can be more prevalent if the bones have been trampled onto a hard or stony substrate, but these breakages are not sufficiently diagnostic to be distinguished from other forms of primary taphonomic fragmentation (e.g. mastication) and this is especially true for fossilised bones (Olsen & Shipman, 1988).

In recent years actualistic experiments involving trampling of modern bone material onto a variety of substrates, including silt (Fiorillo, 1987), coarse sand (Fiorillo, 1989; Behrensmeyer *et al.*, 1986), gravel (Fiorillo, 1987, 1989; Behrensmeyer *et al.*, 1986) and a karstic limestone pavement (Andrews & Cook, 1985), have provided useful taphonomic information on accidental bone modification. The bone assemblage studied by Andrews & Cook (1985) was trampled upon a hard

karstic limestone terrain, with limited soil cover (although there was substantial undergrowth) and they found that dispersal of the bones (from a single skeleton) was extremely rapid within the first six months after death. This dispersal was caused mainly by accidental kicking and trampling, with some gravitational movement down a gully slope into a phreatic tube system. Haynes (1991) also studied the fragmentation and destruction of large mammalian skeletal elements by trampling, whilst Andrews (1990) trampled out owl pellets sealed within plastic bags, to assess the amount of disintegration of the pellets and the fragmentation of larger bones. He noted that whilst still encased in pellets, bones tend to be protected from further modification, but following disintegration they tended to fragment quite badly, although isolated teeth remained intact and identifiable.

During these studies and separate investigations of comparable fossil vertebrate remains, these authors noticed characteristic superficial marks upon the trampled bones. These marks, collectively known as 'trample marks' or 'trample scratches' can be seen with the naked eye and usually comprise fine sub-parallel scratches or scrapes, shallow and confined only to the outer surfaces of remains. They are often closely spaced, they may intersect one another and are roughly linear with various curvature along their length (Andrews & Cook, 1985). Trample marks occur on bones lying on a substrate or partially buried within it and are caused by individual or clustered grains of sand, acting as a "miniature tool leaving V-shaped straie on the bone surface they are ground against" (Fiorillo, 1987, p. 73). They often seem to be more pronounced upon the upturned surfaces of bone, where sand grains were pushed across the surface by pressure of the moving foot (Behrensmeyer *et al.*, 1986).

The amount and type of damage depends also upon the shape and robustness of the skeletal element, for instance spherical specimens such as vertebra centra are less prone to damage, whereas cylinders (e.g. limb bones) are more likely to be covered in scratches. Trample marks also tend to be located on the shafts of long bones, trending roughly transverse across the shafts of linear bones and concentrated toward the centre of curvature of ribs, orientated perpendicular or oblique to the long axis (Behrensmeyer *et al.*, 1986). Within archaeological accumulations trample scratches have previously been confused with butchering scrapes and cut-marks. However, they can be distinguished by their multi-directional orientations (Andrews & Cook, 1985; Olsen & Shipman, 1988).

Modern trample marks have been studied in detail macroscopically and microscopically by several authors (e.g. Andrews & Cook, 1985; Behrensmeyer *et al.*, 1986), in an effort to differentiate them from deliberately-made butchering marks. Andrews & Cook (1985) examined annually the taphonomic history of a cow carcass over a period of 71/2 years and described the alterations made to the bone assemblage by limited scavenging and substantial accidental biological modification by other cattle.

They described six main micromorphologies based upon their study of the skeleton which was trampled upon a hard limestone substrate, similar to that surrounding the Hornsleasow doline. Their types included multiple fine striations (only seen at magnifications greater than x75, and have V-shaped cross-sections); irregular sided flat-bottomed troughs; shallow irregular marks; multiple groups of parallel V-shaped scratches; grooves with multiple, parallel internal striations (seen only at high magnification); and wide U-shaped scrapes composed of multiple striations (Andrews & Cook, 1985, p.685). All were attributed to trampling on a rocky surface, although fine striations and superficial irregular marks may not preserve well in fossil bones, and the large scale scrapes and V-shaped scratches might be easily confused with cut-marks in archaeological sites. Their descriptions are also useful in distinguishing genuine trample marks from scrapes and scratches made upon fossil bones during preparation.

As buried bones are less likely to be susceptible to kicking or trampling damage, an absence of trample marks might suggest rapid burial. However, multiple scoring marks can also be caused after burial by soil movement (Wood & Johnson, 1978). Silty or muddy substrates would preclude abrasion and marking by sand grains (Behrensmeyer *et al.*, 1986), although animals could carry isolated clasts upon their feet and scratches could also result from contact with nails, hooves or claws. Trampling may even cause abrasive polishing to a bone surface in fine-grained sediments (Shipman *et al.*, 1984; Behrensmeyer *et al.*, 1989). Trample marks have been observed in the vertebrate fossil record from the Upper Cretaceous (Fiorillo, 1987) and Tertiary (Fiorillo, 1984, 1987, 1988b,c; Behrensmeyer *et al.*, 1986).

In determining the amount of damage caused by accidental biological modification in the Hornsleasow assemblage, I have assessed presence or absence of trample marks and the percentage area covered with trample marks. Scratches usually occur in sets (Fiorillo, 1984, 1987), and therefore sets of scratches were counted and the percentage covered worked out (for the external surface only).

Around 30% of the cetiosaur bones showed fine surface scratching, comprising shallow roughly linear, sub-parallel sets (1mm wide by 10-20mm long: Fig. 10.11a). This proportion of bones modified is similar to those seen in the Campanian (Upper Cretaceous) bone assemblages reported by Fiorillo (1987), but is much less than other assemblages were trample marks can cover up to 50% of the remains (e.g. Behrensmeyer *et al.*, 1986). Fiorillo (1987) speculated that the difference was caused by the fine grained nature of sediment in the Campanian deposit and this would seem to be true also for the Hornsleasow deposit. The marks observed are not preparation scratches as they can be seen to continue under surface mineralisation crusts of iron and manganese oxides (Metcalf, 1993). The scratches run roughly perpendicular or at an angle to the main axis of the bone, and were seen especially well upon the flat surface or shafts of ribs (Fig. 10.11a), a jaw fragment, vertebral processes and the ilia. These

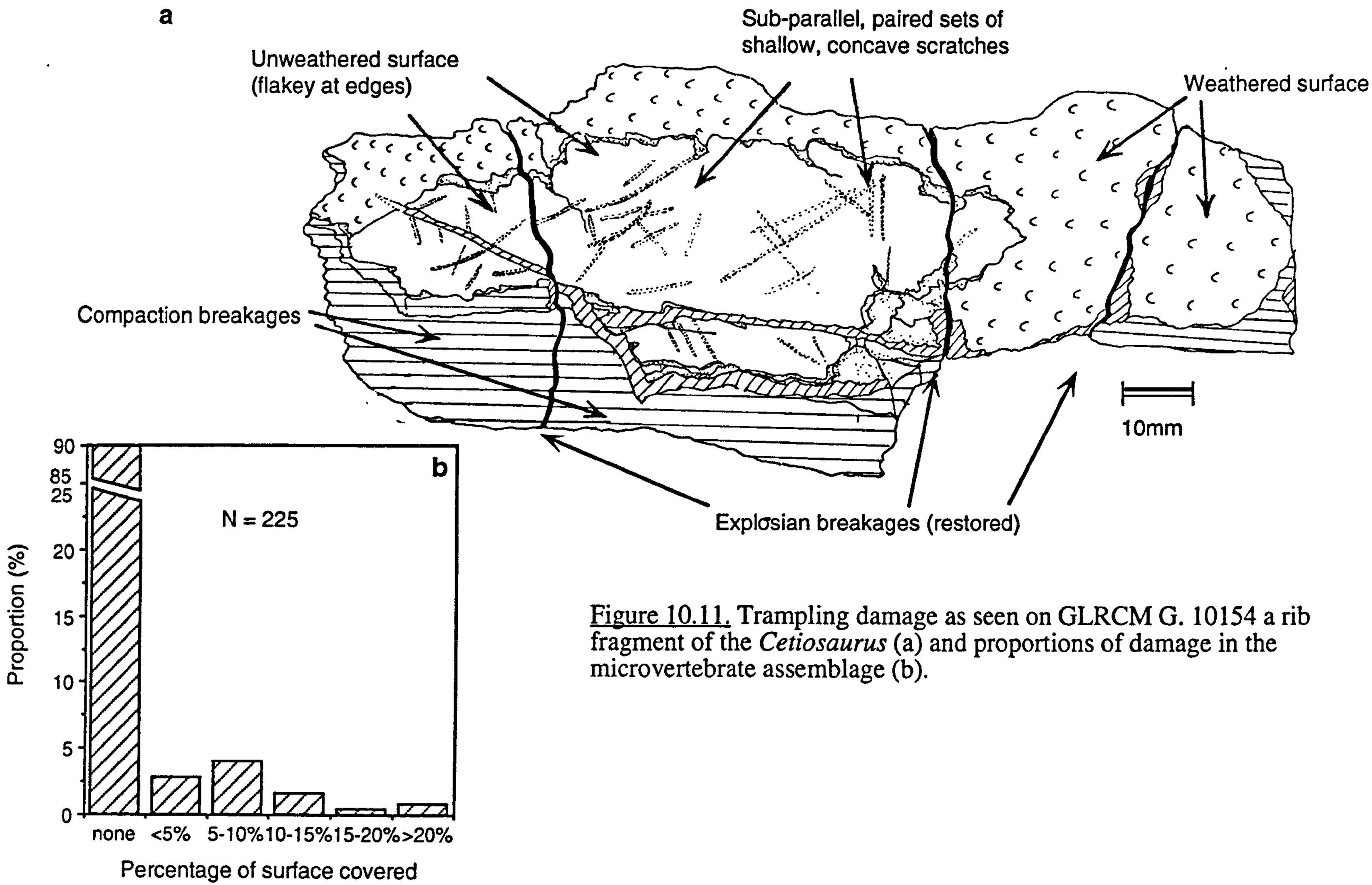


Figure 10.11. Trampling damage as seen on GLRCM G. 10154 a rib fragment of the *Cetiosaurus* (a) and proportions of damage in the microvertebrate assemblage (b).

scratches are similar to those described above and are attributed to trampling by contemporary animals. Some of the spiral fracturing seen on the bones may have been caused by accidental biological modification whilst the bones were still fresh. The trample marked surfaces indicates that the bones were exposed for a considerable amount of time (Fiorillo, 1988b).

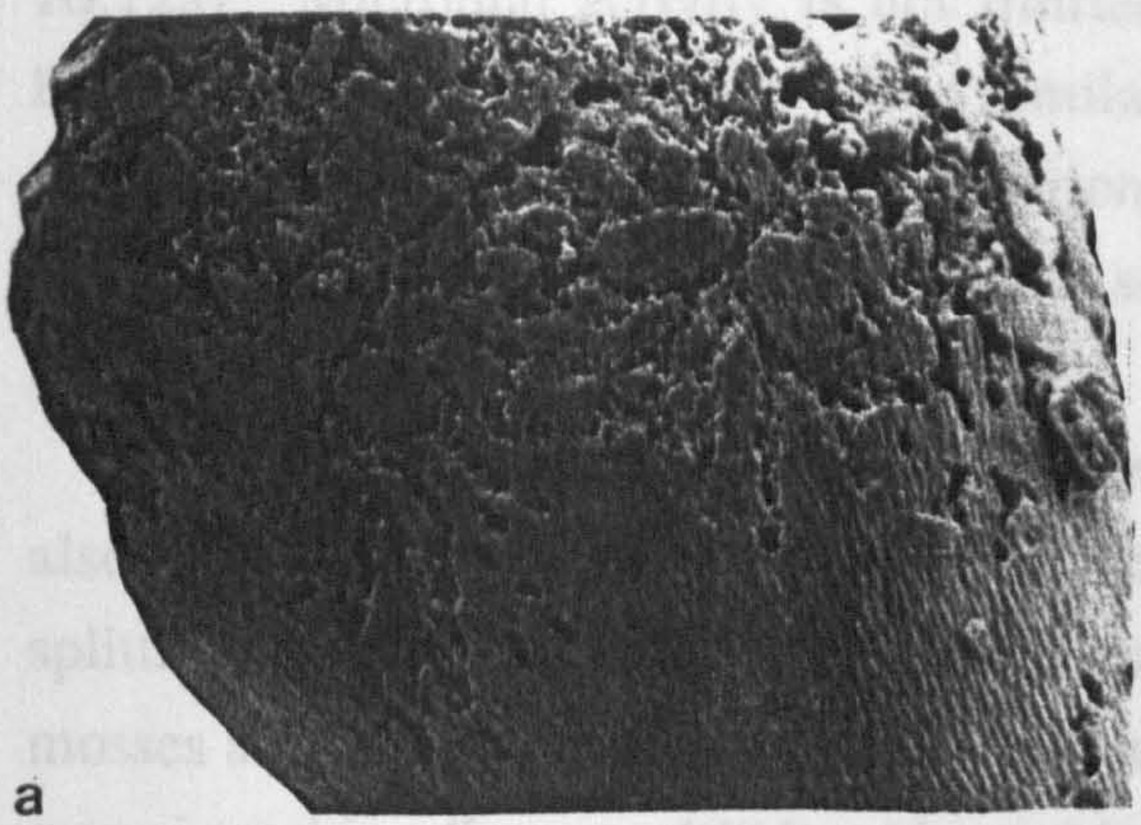
In all 90% of bones in the microvertebrate assemblage are free of any sign of trampling marks, and of the 10% which show scratches, over half only show 5% of their surface covered (Fig. 10.11b). This low frequency could be mainly due to lack of recognition on my part, as trample marks are unlikely to be distinguishable upon such small specimens. Also their small size means that multiple scratches and scrapes (the most easily recognised trample marks) are unlikely to occur. It is probable that the remains were trampled into the sediment as much as the larger bones, but their small size would have meant that they would quickly become buried. It is possible that the soil-forming processes, and the churning action of animal activity around the pool, served not only to separate and fragment the remains, but to bury and hence, preserve them.

10.4.2. Biological modification of buried vertebrate remains

Phosphate is an essential nutrient for most organisms, but is rare in most environments (Martill, 1991), therefore a number of vertebrate and invertebrate animals feed upon bone which has lost its organic matter. Scavenging and bone utilisation by vertebrates has already been considered in section 10.3., but other organisms are also known to make use of bone material, and some of their modification is recognised in Mesozoic fossil assemblages. In the marine environment, a number of invertebrates such as gastropods and echinoderms are known to feed upon bone phosphate, however, fresh-water gastropods such as those recovered from the Hornsleasow accumulation do not. Certain terrestrial insects, such as termites, and moth and beetle larvae, are known to chew bone material, leaving behind characteristic rounded bore holes or tubes (Behrensmeyer, 1975; Shipman, 1981). Although beetles are a common component of Middle Jurassic arthropod assemblages, no such distinctive marks were found upon the Hornsleasow bones and teeth.

A number of micro-organisms, such as fungi and bacteria, are also known to utilise bone in modern marine environments and microbial decay is one of the most important factors leading to the destruction of skeletal material (Martill, 1991). Microbial utilisation weakens bones so that they become susceptible to mechanical breakage and typical dendritic borings made by such activity has been recognised upon vertebrate material (mainly fish and teleosaur crocodile) from Middle Jurassic marine assemblages (e.g. the Callovian Oxford Clay, Martill, 1989, and the Bathonian Charlbury and Taynton Formations, pers. obs., and Metcalf & Underwood, 1995: Fig.

Figure 10.12. Acid digestion and secondary biological erosion. Microbial borings: (a) H1.1., fish tooth from the marine Charlbury Formation (Bathonian) at Brockhill Quarry, Gloucestershire. Acid erosion associated with root or humic acids in the soil: (b) G. 58531, split crocodile tooth, showing pock-marking upon the exposed breakage surface through the dentine ('dn') (field of view = 740µm); (c-e) G.66768, theropod tooth showing incomplete stripping, as well as pitting and pockmarking of enamel and underlting dentine: (c) lateral view of whole tooth showing distribution of enamel (unshaded) and dentine (stippled), (scale bar = 1mm); (d) inset to (c) showing close up of pitted enamel ('el') overlying deeply pock-marked dentine (field of view = 890µm); (e) view of distal carina and edge of tooth, showing pock-marked enamel and complete removal of the enamel beads overlying carina serrations (field of view = 1.9mm). Preparation acid digestion: (f) Fine enamel pitting at tip of crocodile tooth immersed in 5% acetic acid (unspent and unbuffered) for 2 hours (field of view = 200µm).



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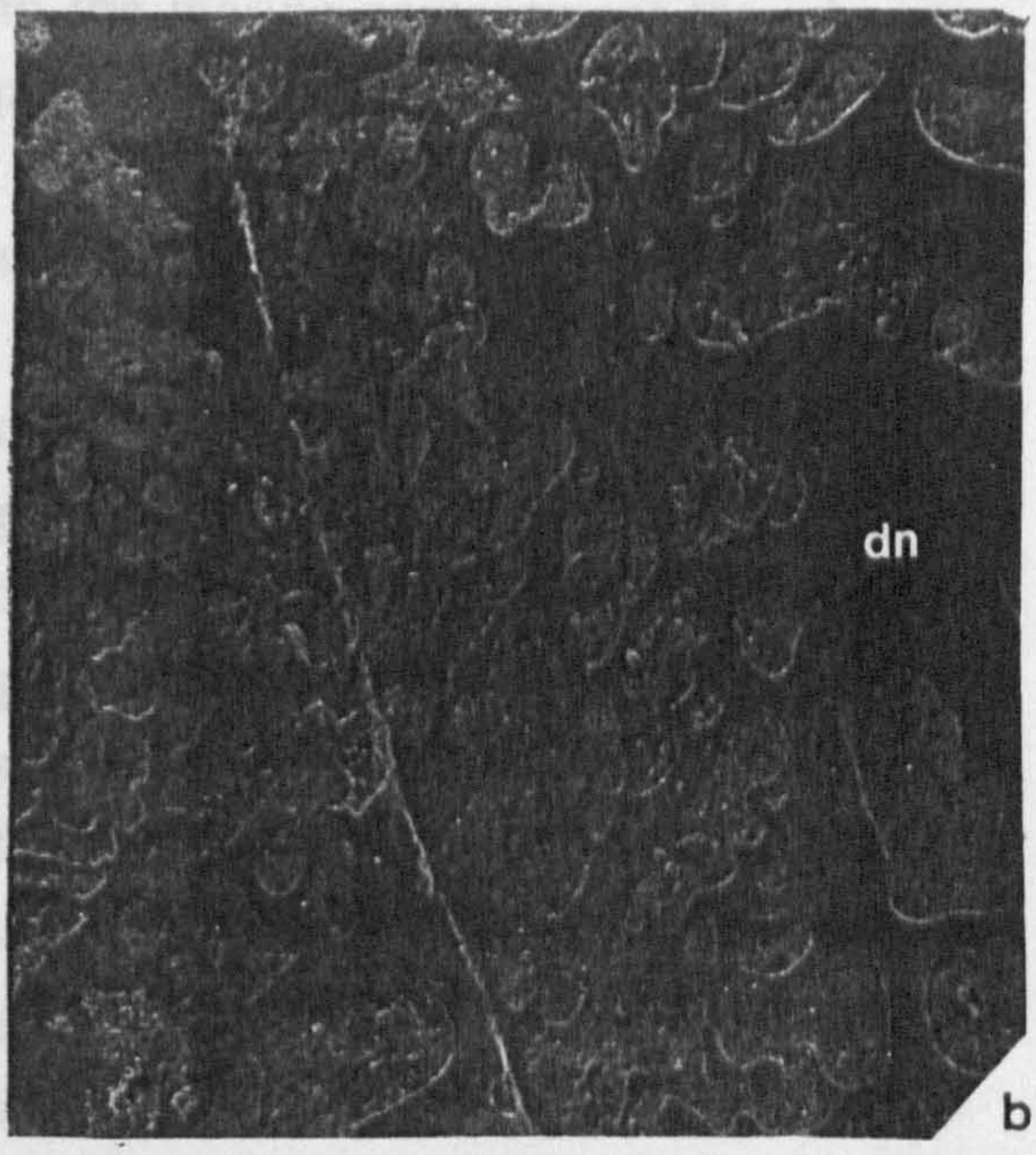

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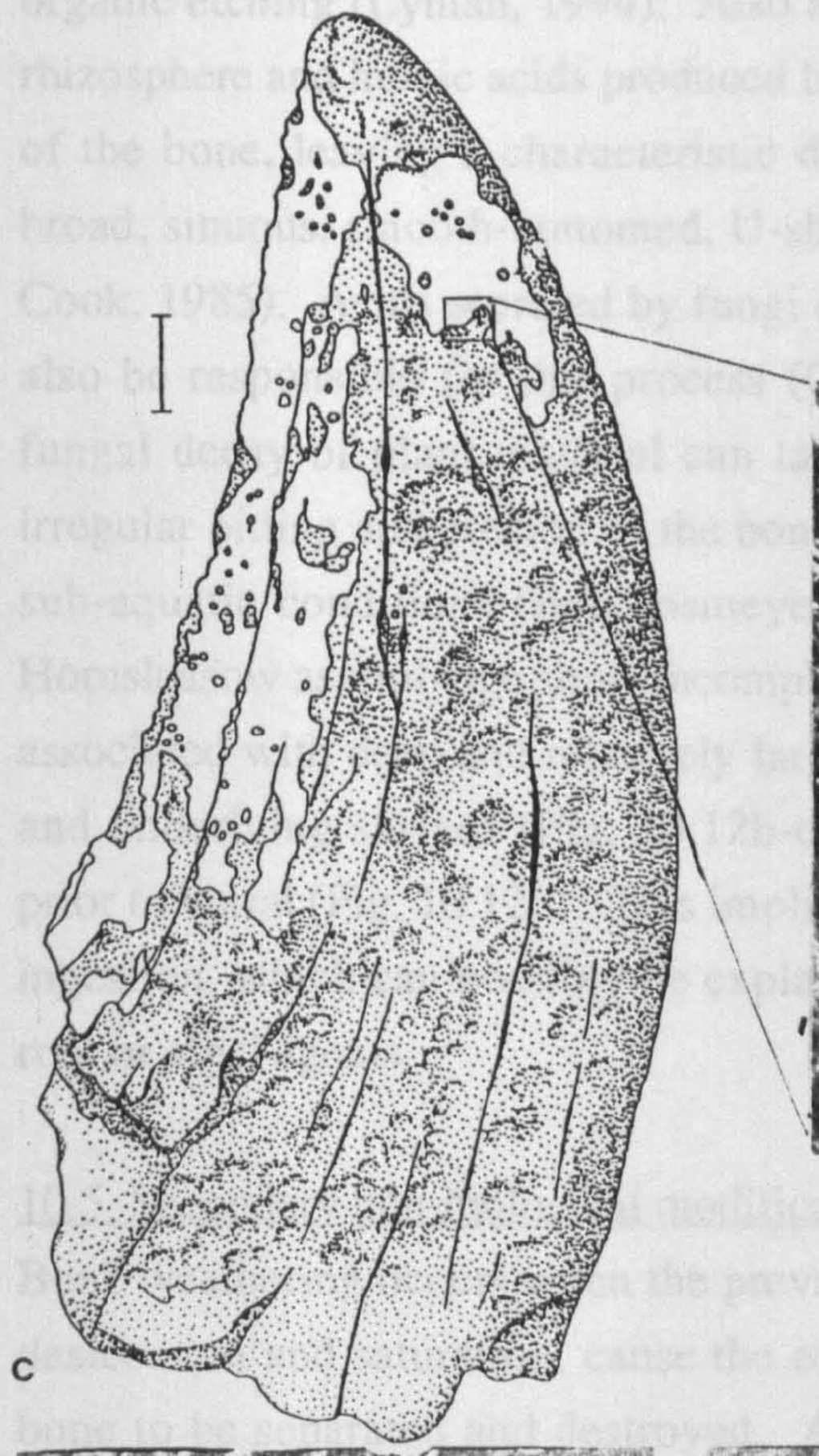

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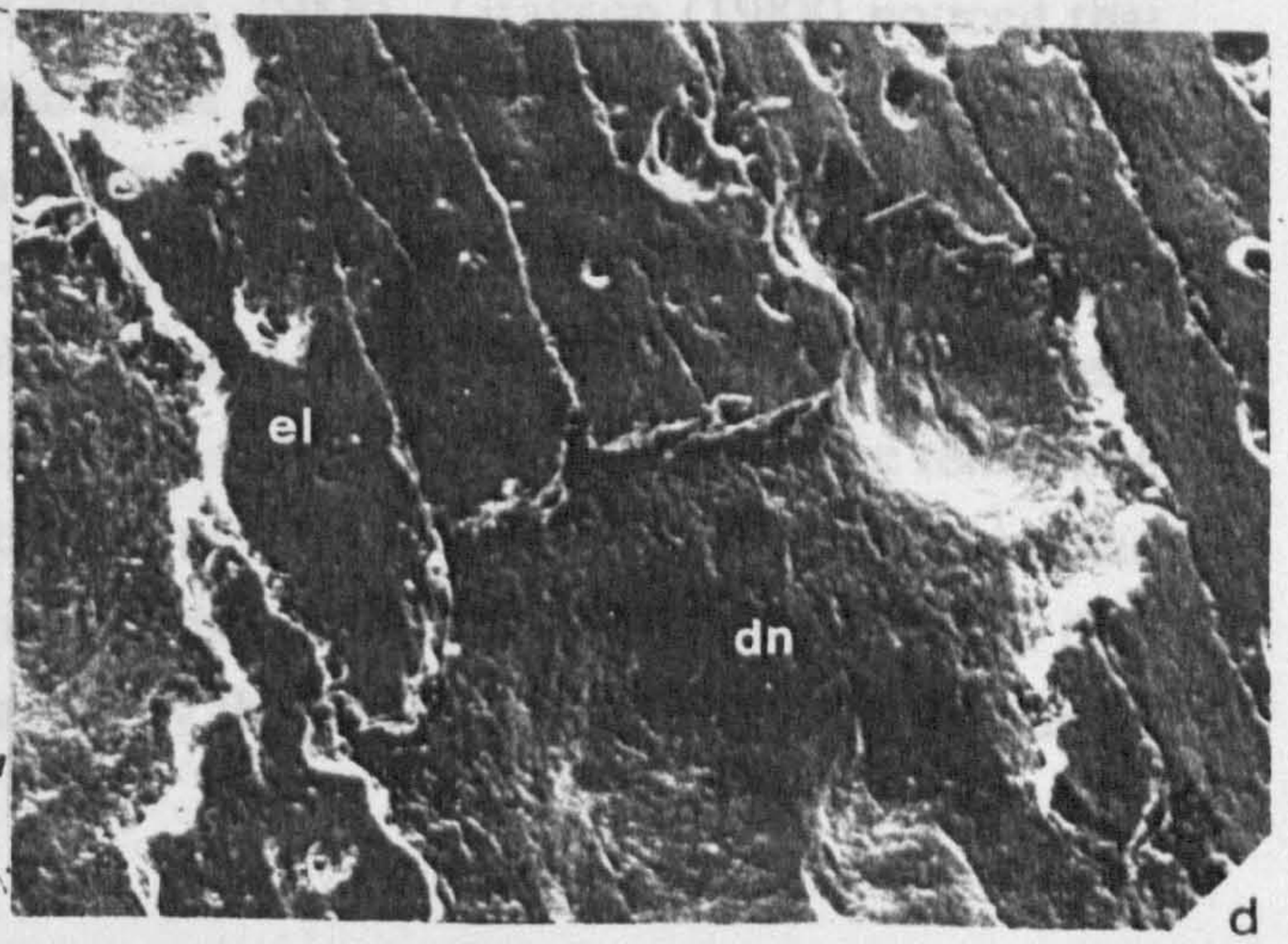

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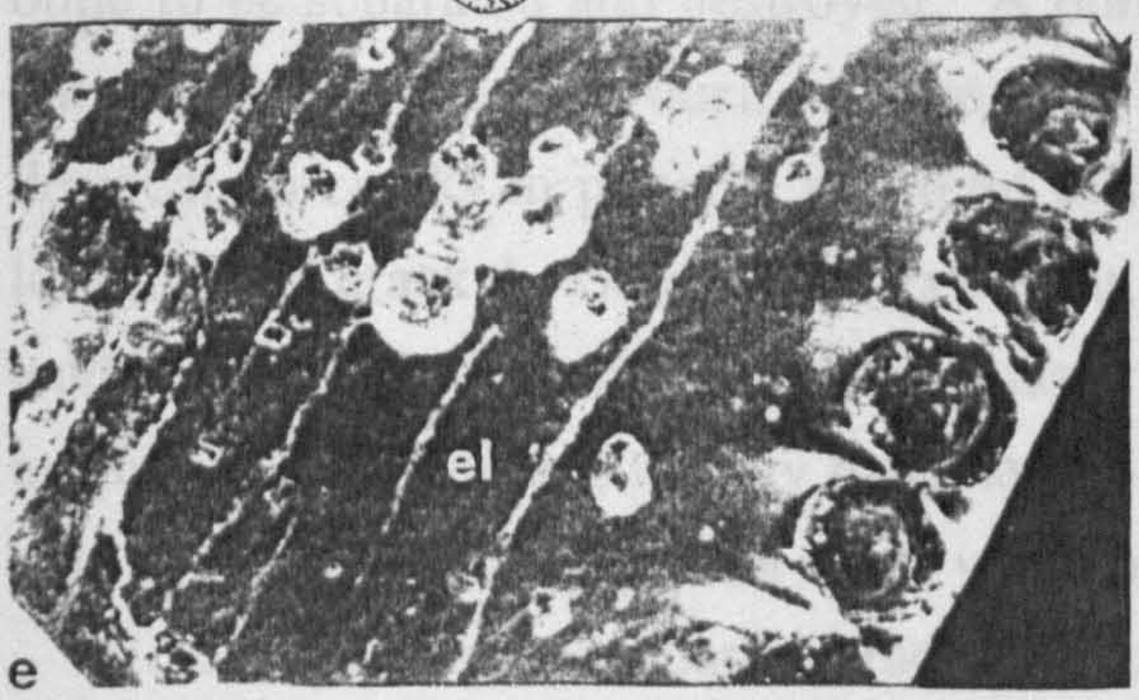

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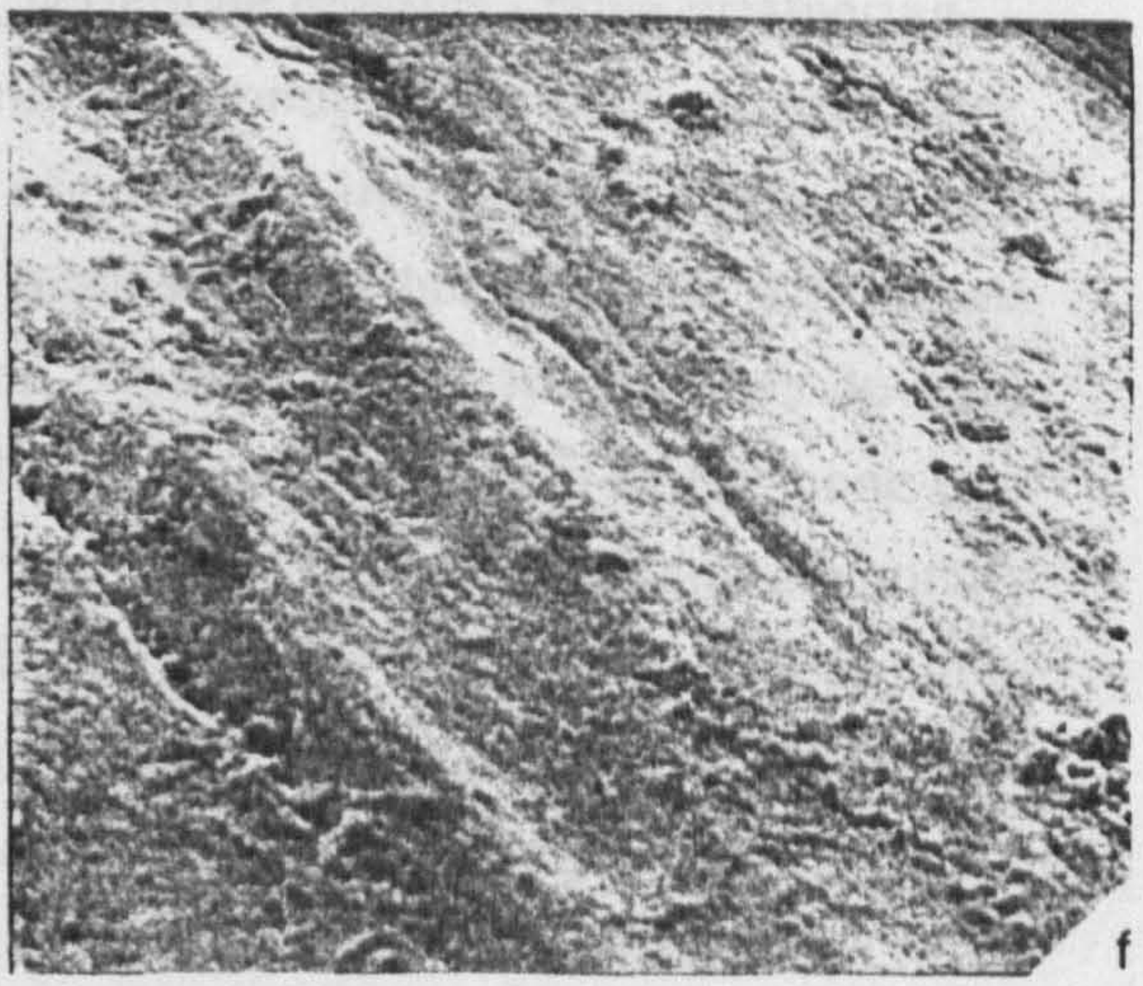

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10.12a). Microbial activity is not limited to the marine environment though, as Behrensmeyer (1978, 1991) noticed similar patterns within weathered bones on the surface and within soils in Amboseli National Park, Kenya and within freshwater sub-aquatic environments. However, this sort of biological modification was not recognised in the Hornsleasow assemblage.

Plant roots and decaying plant matter contained in the humic layer of soils are also known to modify bone material. The growth of roots may be responsible for splitting and breaking bone material on the surface or shallow buried and the growth of mosses and lichens upon bone material occurs prior to burial and can result in pre-burial organic etching (Lyman, 1994). Also acids associated with the growth of roots in the rhizosphere and humic acids produced by the decay of organics can dissolve away parts of the bone, leaving a characteristic dendritic appearence (Behrensmeyer, 1978) or broad, sinuous, smooth-bottomed, U-shaped grooves etched by root acid (Andrews & Cook, 1985). Acids secreted by fungi associated with decomposing plant matter may also be responsible for this process (Grayson, 1988). Grayson (1988) noticed that fungal decay of plant material can take place in relatively dry conditions. Rough irregular pitting and erosion of the bone surface can also be caused by algal growth in sub-aquatic conditions (Behrensmeyer, 1991). One or two teeth and bones in the Hornsleasow assemblage show incomplete stripping of the enamel or outer bone layers, associated with deep and relatively large pock-marks developed within the remaining and underlying surface (Fig. 10.12b-e). Some of this occurs upon surfaces broken prior to burial (Fig. 10.12b). This implies a form of selective acid digestion, rather than ingestion, which can possibly be explained by the localised acid erosion from humic, root or algal agents.

10.5. Secondary non-biological modification of vertebrate remains - bone weathering

Bone weathering occurs when the prevailing climatic conditions, such as temperature, desiccation and saturation, cause the original organic and inorganic components of a bone to be separated and destroyed. A bone begins to weather once the soft tissues which surround it have been removed. When exposed to air, fresh bones develop desiccation cracks in patterns that tend to follow the original grain of the bone (i.e. longitudinal cracks). Given more exposure, these cracks will multiply and eventually lead to total disintegration of the bone tissue (Behrensmeyer, 1991). Bones exposed to bright sunshine become rapidly bleached. This leads to the loss of organic matrix and may result in the disintegration into a white powdery substance (Martill, 1991).

Longitudinal desiccation cracks are usually fairly shallow cracks confined to the surface of the bone (Fig. 10.13i; Behrensmeyer, 1978; Hill, 1980; Fiorillo, 1988b,c), and can usually be seen with the naked eye. Chemical weathering begins at the bone surface, eventually progressing into the interior of the bone. In severe cases,

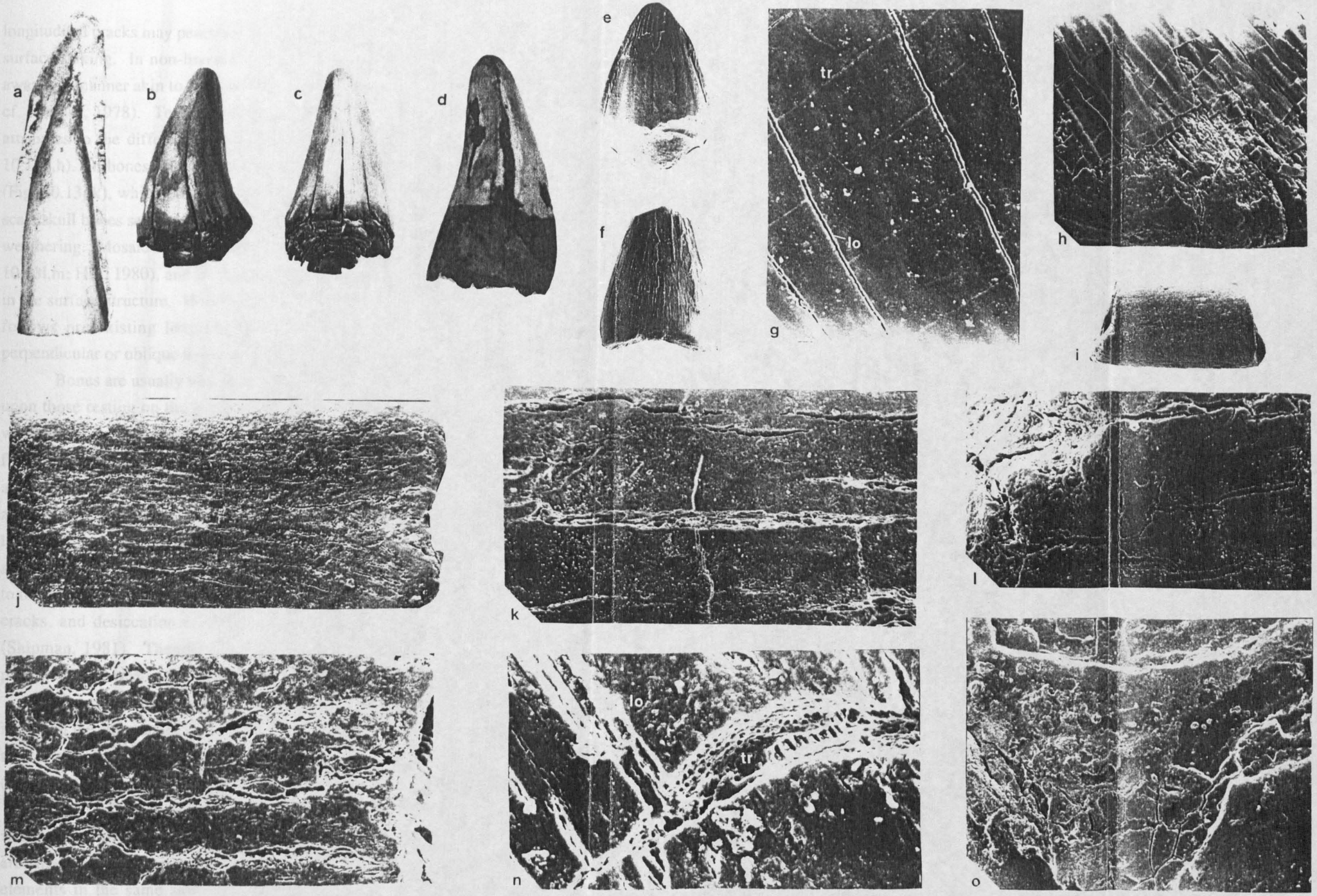


Figure 10.13. Weathering features on vertebrate material. Modern crocodile teeth (a-d) showing: (a) complete separation along deeply penetrating longitudinal crack (stage '3') (crown length = 21mm); (b-c) enamel stripping, dentine penetrating longitudinal cracks and onion-skin exfoliation pattern of dentine layers (stage '3') (crown length = 40mm); (d) and flaking of enamel (stage '2') (crown length = 34mm). Fossil teeth (e-h) showing: (e) G. 52004, enamel stripped crocodile tooth (stage '2') (crown length = 3.2mm); (f) G. 56906, cracked and enamel flaked tooth (stage '1'('2')) (crown length = 1.4mm); (g) G.61411, longitudinal cracks ('lo') and fine transverse splits ('tr') in enamel of crocodile tooth (field of view = 120μm); (h) G. 61411, flaking of enamel at base of tooth (field of view = 240μm). Fossil bones (i-n) showing: (i) G. 56918, limb-bone showing fine longitudinal splits (length = 1.7mm); (j) G. 56917, bone fragment, showing longitudinal splits (length = 1mm); (k) G. 56914, lense-shaped longitudinal split with radiating transverse cracks in bone fragment (field of view = 380μm); (l) G. 56918, mosaic cracking in bone fragment (field of view = 320μm); (m) G. 61609, irregular splitting in bone fragment (field of view = 360μm); (n) G. 61410, deep longitudinal and transverse fractures (field of view = 100μm). Fossil fish scale (o) G. 51450, cracking and flaking of enameloid layers (field of view = 0.8mm).

longitudinal cracks may penetrate the interior cortex of a bone, and are accompanied by surface flaking. In non-linear bones (e.g. vertebrae) and teeth the surface may peel away in a manner akin to the onion-skin weathering of granite boulders (Fig. 10.13b,c; cf. Holmes, 1978). Teeth show fine splitting in the enamel which Andrews (1990) attributes to the differential contraction of enamel and dentine to desiccation (Fig. 10.13g,h). In bones, microscopic lens-shaped splits develop between collagen fibres (Fig. 10.13j,k), which later develop into penetrative longitudinal cracks. On a larger scale skull bones separate along sutures and teeth fall out of sockets during progressive weathering. Mosaic or tessellated cracking may also occur upon articular facets (Fig. 10.13l,m; Hill, 1980), and in tooth enamel (pers. obs.), and is formed by irregularities in the surface structure. However a similar pattern is seen when transverse fracturing follows pre-existing longitudinal crack-lines as well as new lines of weakness perpendicular or oblique to the bone fabric (Figs. 10.6a,h & 10.13h,n,o).

Bones are usually weathered more on the exposed upward-facing surfaces than upon those resting on the ground. Buried bones are thought to be less susceptible to weathering, and microvertebrate remains enclosed in faecal pellets are also protected from weathering (Andrews, 1990). However, desiccation cracks can also be formed after burial by post-burial or diagenetic processes (Potts, 1986; Lyman & Fox, 1989) and hence, bones may continue to weather in subsurface conditions. However, very little work has been completed to distinguish between subaerial and surface weathering, although buried bones are thought to weather at a much slower rate than those exposed to the elements. Transverse fractures may follow and enlarge pre-existing longitudinal cracks, and desiccation cracks may also be found in experimentally heated bones (Shipman, 1981). Therefore caution is stressed when ascertaining the amount and pattern of bone weathering from surface features in a fossil bone assemblage.

The rates and degree of bone weathering are controlled by several factors, including the structure of the bone itself, the prevailing climatic conditions surrounding the bone and the length of time the bone was exposed to these conditions. Lyman (1994) outlined three specific controls:

(1) Bones of different shape, size and structure exhibit different rates of weathering, for instance Behrensmeyer (1978) saw that phalanges and podials, which are small and structurally compact bones, weathered much more slowly than other elements in the same skeleton. However, little is known on the relative rates of disintegration for different skeletal elements.

(2) Bones of different taxa, and especially those of different original body size, weather at different rates (Behrensmeyer, 1978; Gifford, 1981). Again this is probably controlled by differences in bone structure, porosity and density.

(3) The surrounding climatic conditions, in particular moisture levels and temperature, play a part in rate of bone weathering, and especially if they tend to

fluctuate. For example, Behrensmeyer (1978) found that bones advanced to higher states of weathering in an open plain environment, than in forested areas or wetlands in an arid climate. Microenvironmental parameters such as vegetation cover, localised moisture levels, substrate and association with other carcasses may be more important in bone weathering than overall prevailing weather conditions (Behrensmeyer, 1978). However, Andrews & Cook (1985) found no evidence of weathering upon partially buried, vegetation covered and exposed bones of a cow carcass examined after 71/2 years exposure in a temperate, wet environment.

The progressive nature of bone weathering has led to a scale of weathering stages to be established, which for some environments indicate how long a bone was exposed prior to burial (Table 10.1. after Behrensmeyer, 1978; Lyman & Fox, 1989). Behrensmeyer's scale was developed upon macroscopic features of bone weathering seen upon a $1cm^2$ area of the surface, for animals over 5kg in body weight. She studied various skeletal assemblages over a period of 15 years in Amboseli National Park, Kenya which has a semi-arid environment. The scale is a continuous spectrum with arbitrary divisions called 'weathering stages' set by the observations made by Behrensmeyer (1978: Table 10.1). Behrensmeyer (1978) found that time spans corresponded with degree of weathering and she suggested that this might be true for fossil bone assemblages. Exposure times of only a few months are sufficient for stage '1' surface cracking to develop in arid conditions (Table 10.1: Behrensmeyer, 1978). Modern crocodile teeth showing weathering stages between '2'-'3' (Figure 10.13a-d) were picked up from surface exposures in an arid climate (Kenya) during the summer of 1992 by Mr. N. Rutger and he estimated that they had laid on the surface for less than a year (pers. comm.). Extremely weathered bone material, in Behrensmeyer's weathering stage 5, indicate anything from 6-15 years of exposure (Table 10.1). However, bones may advance between two or three stages relatively quickly, then remain in another for a considerable length of time (Gifford, 1981), so that the scale is only a rough guideline, providing the best estimate of the minimum duration since initial exposure. Time spans could be worked out for the Amboseli Park fauna as Behrensmeyer knew how long the skeletons had been exposed when she was examining their weathering stages.

The stage of weathering displayed by a bone indicates the rate and duration of weathering on that specimen (Lyman, 1994). Therefore the weathering profile for an assemblage should show the rate and degree of weathering for the whole accumulation. However, fossil bone accumulations consist of specimens of various skeletal elements from many skeletons of several different taxa. These may have died at different times, and the bones even within a single skeleton may have been exposed for different lengths of time. There is the added problem that in an attritional assemblage weathered bones may be introduced into the accumulation from other regions. One way to solve these

(a) Large modern mammal bones			(b) Small modern mammal bones		(c) Fossil bones	
Weathering stage	Description	Exposure duration	Description	Exposure duration	Weathering stage	Description
'0'	Bone surface fresh, greasy, perhaps with skin or ligament/soft tissue attached. No cracking.	0-1 years	No modification	0-2 years	'0'	Bone surface fresh; no sign of flaking or cracking.
'1'	Surface splits appear, usually parallel to the fibrous grain of the bone - 'longitudinal cracks'. Articular surfaces with mosaic cracking of covering tissue and bone	0-3 years	Slight lenticular splitting of bone parallel to fibrous structure; splits develop in enamel and dentine layers in teeth.	1-5 years	'1'	Surface longitudinal cracked. Cracks confined to the outermost layers of bone.
'2'	Flaking of outer surface (exfoliation), straight-sided cracks appear. Marrow decays.	2-6 years	More extensive splitting, little flaking. Flaking of tooth enamel and exfoliation of dentine layers.	3-5+ years	'2'	Surface flaking (exfoliation), tesselated (mosaic) pattern of cracks. Cracking begins to penetrate into bone cavities.
'3'	Roughly homogeneous altered compact bone resulting in fibrous texture; cracks begin to penetrate 1-1.5mm depth.	4-15 years	Loss of deep flakes between splits; extensive splitting and exfoliation of dentine.	4-5+ years	'3'	Extreme outermost layers of bone gone (onion-skin exfoliation). Cracks penetrate into bone cavities; microvertebrate remains may be split open.
'4'	Coarsely roughened fibrous surface; loose splinters of bone on surface; open cracks penetrate into inner cavities.	6-15 years				
'5'	Bone disintegrating *in situ*, large splinters present, bone material very fragile.	6-15 years				

Table 10.1. Weathering stages in (a) large modern mammalian skeletons (after Behrensmeyer, 1978), (b) modern micromammalian assemblages (after Andrews, 1990), and (c) fossil bones (after Fiorillo, 1988a).

problems is to concentrate upon analysing only one particular elemental type from a single species (Lyman & Fox, 1989) or if possible all the elements from a single skeleton (Lyman, 1994). The second approach was used for the cetiosaur remains, which showed a narrow span of weathering typical of a single skeleton recovered from one stratigraphic horizon (i.e. not reworked).

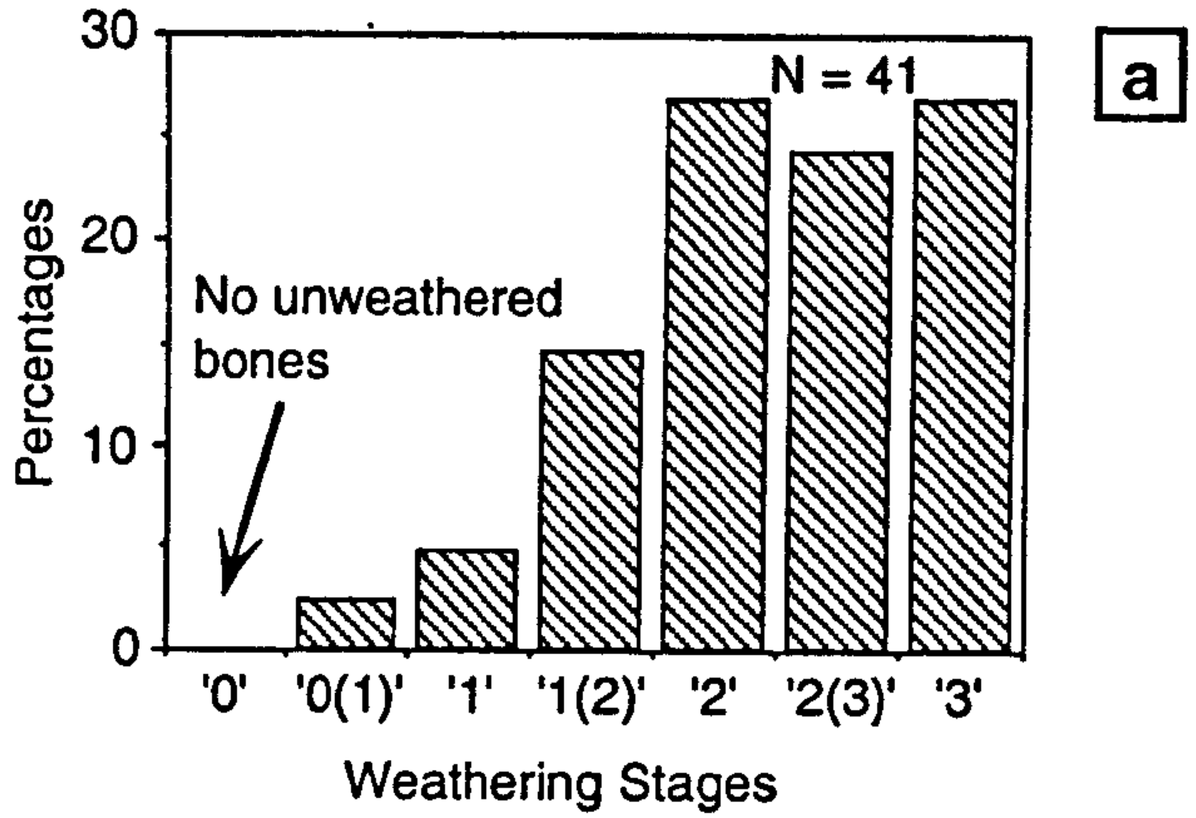
Variable degrees of weathering for an assemblage, such as those exhibited by the microvertebrate assemblage, have been used as evidence for attritional accumulation (Potts, 1986), but microenvironmental conditions such as variation in vegetation cover can also result in bones showing different weathering stages (Lyman & Fox, 1989). This can be tempered by assuming all specimens in an assemblage weathered in the same microenvironment.

Behrensmeyer's stages have formed a solid basis to explore the various weathering processes, their time spans and effects, and other authors have since improved on them. In 1990, Andrews examined weathering processes which effect micromammalian bones and compared his weathering stages with those of Behrensmeyer (1978) and these are summarised in Table 10.1 (with additional observations of my own). Fiorillo (1988b,c) also developed a simplified scale for fossil bones based on his interpretation of Behrensmeyer's work (Table 10.1), which I consider to be more relevant to this study and this is the one used in these analyses.

In this study I examined the whole surface of the bone (rather than the $1cm^2$ suggested by Behrensmeyer, 1978), under direct lighting (for the macrovertebrate remains) and with the aid of a binocular microscope for the microvertebrates. Microscopic weathering features seen on the microvertebrate material, such as lenticular desiccation splits and cracked enamel were later examined and photographed under the SEM (Fig. 10.13). For each specimen, the degree of weathering was assessed and a weathering profile produced, showing the percentages of bones in the assemblage within each category (Fig. 10.14). In this study, weathering stages were interpreted on the assumption that subsurface alteration was minimal.

10.5.1. Weathering stages for the cetiosaur skeleton

The degrees of longitudinal cracking and flaking of the surface was observed for 41 bones of the cetiosaur skeleton, and assessed in terms of Fiorillo's (1988b) scale of weathering stages. Although weathering is habitat and climate dependent, the conditions within the limestone hollow at Hornsleasow during the period of exposure (prior to paleosol development) may have been sufficiently comparable to the semi-arid climate to that of the Kenyan examples, so that the bones were studied with little change to the scheme adopted by Behrensmeyer (1978) and adapted for fossil assemblages by Fiorillo (1988b,c).



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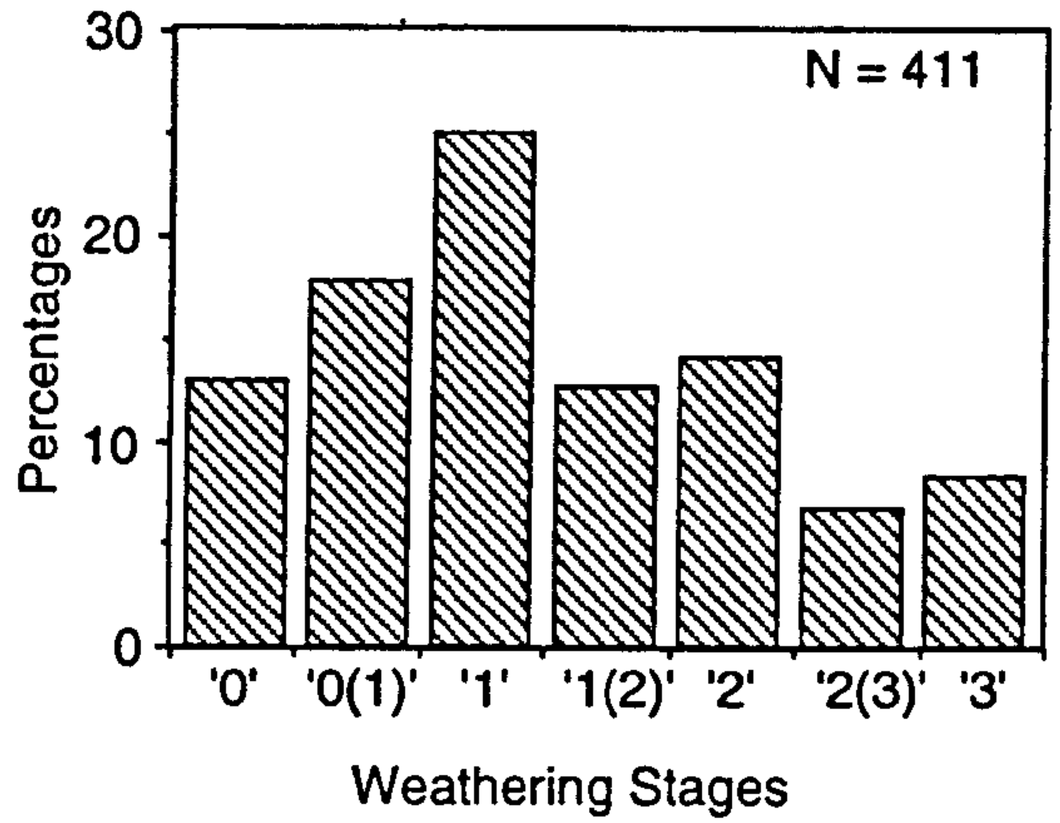

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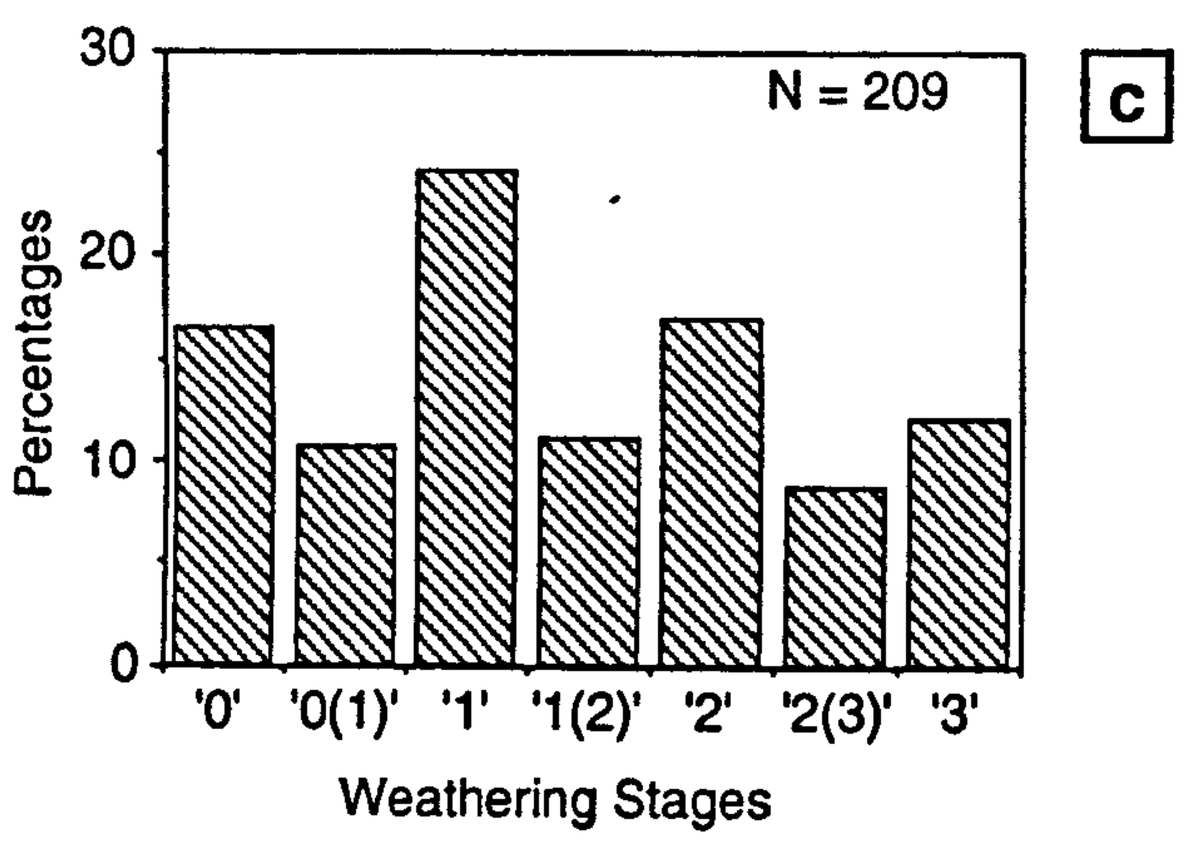

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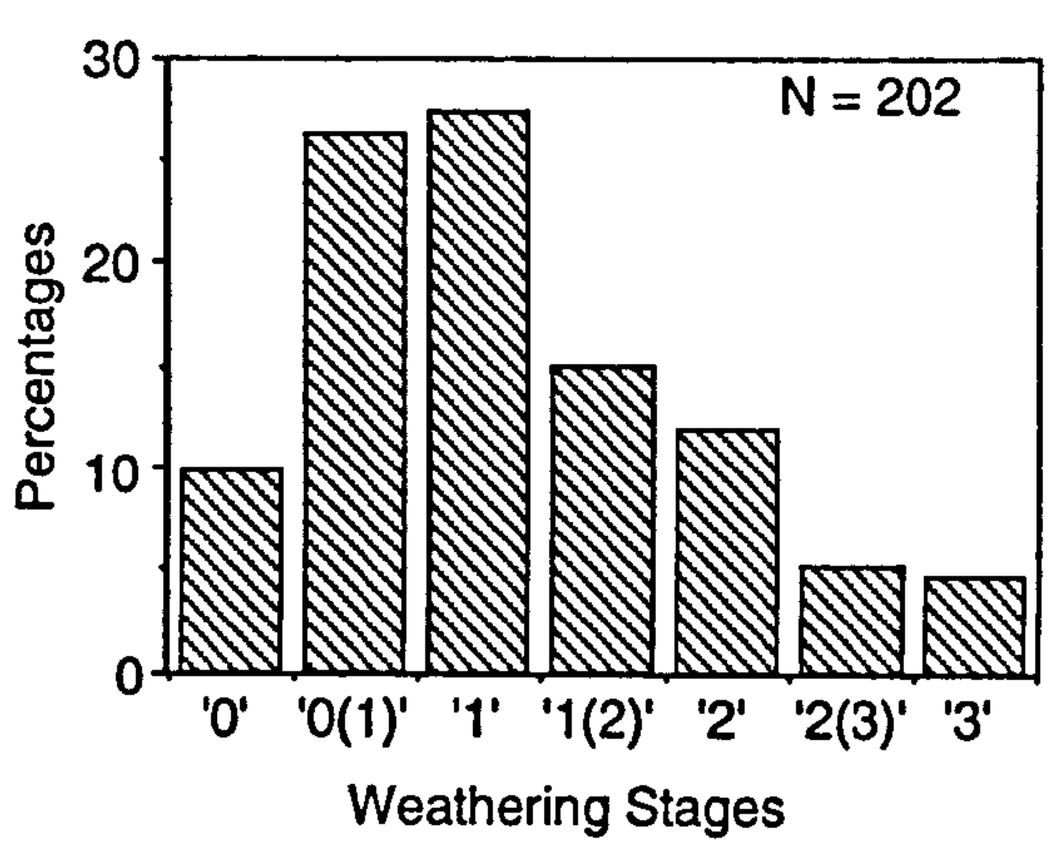


d

Figure 10.14. The weathering profiles for (a) the macrovertebrate assemblage and (b) the microvertebrate assemblage. The microvertebrate weathering profiles were further subdivided into graphs showing the results for bones, dermal armour and fish scales (c) and teeth (d). The number (N) of specimens is indicated. Categories in parenthesis indicate a bone showing traces of a higher category.

The results of the weathering stage analysis are shown in Fig. 10.14a, and illustrated by the ilium and vertebrae in Fig. 10.7. The weathering profiles show that most (78%) of the cetiosaur elements exhibit advanced stages of weathering, with a narrow unimodal peak on the '2'-'3' stages, and there are no unweathered bones. On this profile categories in parenthesis indicate a bone which shows traces of the next higher stage. The narrow range suggests that the cetiosaur skeleton as a whole was exposed for anything from five to fifteen years before final burial. This period is based upon Behrensmeyer's (1978) calculation for the weathering stages of modern bones in arid conditions, but it is considered to be a reasonable minimum estimate of exposure time if the conditions were more wet (bones exposed in swampland in Amboseli National Park weathered at a reduced rate to those in more open locations: Behrensmeyer, 1978). Parts less weathered include the smaller and more compact elements (e.g. rib fragments and detached neural processes) which would have been buried quickly by trampling.

Over half (55%) of the bones including wide, linear elements (limbs, pelves: Fig. 10.7a,b) show differential weathering. The sides and edges of the bones exposed directly to the climate are badly damaged (Fig. 10.7a), whilst the surface lying directly upon the karst and parts of the edges protected by a minimal soil cover are much less weathered (Fig. 10.7b). For clarity the results of the weathering profile (Fig. 10.14a) only show the weathering stages of the more weathered surface. Less stable bones, such as the *Cetiosaurus* neural complex (Fig. 10.7e) would have rolled around in the karstic hollow, and hence, suffered weathering on all sides with equal magnitude. The bone is at weathering stage '3', as the surface remains only within the hollows, and is seen to be exfoliating away from the edges in classic onion-skin weathering.

The depositional environment for the bones is within a karstic hollow with little soil cover; karstic weathering proceeds by chemical dissolution of the limestone by atmospheric and soil carbon dioxide (as carbonic acid) and humic acids (Chapter 4). The friable nature of the Chipping Norton Formation oolites has meant that many small clasts appear within the Hornsleasow deposit, some of these have suffered dissolution upon the karst to produce rounded pebbles. It is possible that bone material could also undergo a limited amount of erosion by chemical weathering, upon the karst and within the soils. The neural complex is in fact a 'bone pebble', and it is likely that this element was rolled or kicked around the base of the hollow for some time before becoming mired and stabilised by soil cover and/or vegetation (Metcalf, 1993).

10.5.2. Weathering damage on the microvertebrate remains

The microvertebrate remains show a much broader span of weathering stages than the larger material (Fig. 10.14b). The wide distribution of bone material weathering stages (Fig. 10.14c) may be a true indication that the microvertebrates are of a mixed

provenance, but again this might also be a function of their small size. The smaller the element, the more easily it is buried, effectively removing it from further modification. Thus, such a bone may not show the true pattern for all the remains of all that animal. If a bone itself displays a broad range of weathering stages, it may have undergone slow burial or re-exposure due to exhumation and reworking (Lyman, 1994). Finally severe weathering (stage '3') will destroy small bones entirely (Behrensmeyer, 1978), giving an under representation of the higher categories of the effects.

The teeth were separated out from the results (Fig. 10.14d) as these more resistant elements would tend to skew the graphs toward the less altered end of the weathering spectrum. Low to moderate weathering is visible upon the teeth (53% in stages '0'-'1'), as the enamel is mostly split, but does not tend to flake unless previously damaged by digestion or breakage. Few teeth showed serious weathering damage (<20%: Fig. 10.14d), or the characteristic exfoliation of enamel and dentine layers seen on arid weathered modern crocodile teeth exposed at the surface for a period of no more than a year and in most cases only a few weeks or months (N. Rutger, pers. comm., 1992; Fig. 10.13a-d).

In less weathered specimens it was possible to observe cracking upon the surface of bones and teeth under the microscope, whilst at higher magnifications (under the SEM), micro-splits can be seen running parallel to the fabric of the bone (Fig. 10.13j). These are usually no more than 200μm in length and less than 10μm in width, they tens to be lenticular or irregular in shape (Fig. 10.13k-m), and often run in closely parallel, but stepped sets across the surface of the bone (Fig. 10.13i,j). They seem to penetrate only the surface of the bone, although at broken edges or epiphyses they can be seen to penetrate deeper, and lower layers seem to have developed them too (Fig. 10.13i,j). The more irregular shaped cracks, seem to develop triple point radiation, finally resulting in a tessellated or mosaic pattern of interlocking cracks (Fig. 10.13l,m). This type of cracking may lead eventually to flaking of the surface between the surrounding splits. It is thought to be a result of irregularities in the bone tissue (Hill, 1980). Some of the splits in the bone surfaces seem to have been enlarged and have regular-sided transverse fractures nucleating from them (Fig. 10.13n). These later fractures seem to be utilising the older cracks, and are easier to see at lower magnifications, so caution should be used when ascertaining the degree of weathering from binocular microscope studies alone. Tooth enamel shows much more regular, straight-sided desiccation cracks running along the length of the tooth. These are usually extremely fine (1-2μm across), but penetrate through the surface layer of enamel, causing flaking in very bad cases (Fig. 10.13e,f,h). Flaking is particularly pronounced were there has been pre-weathering damage, such as acid digestion or at the edge of a breakage. Some also seems to be post diagenetic. Enamel-dentine fossilise at different rates which causes splits to occur. Many of these seem to be typically

transverse to the original desiccation cracks, and less penetrative, but they frequently cause almost square-shaped flakes of enamel to exfoliate (Fig. 10.13h). This sort of flaking is also common in the enamaloid surface of fish scales (Fig. 10.13o).

Lack of longitudinal cracks on bones (stage '0': Fig. 10.14c: 17%) indicates rapid burial or that the remains were protected by water, vegetation cover or tissue until burial (Behrensmeyer, 1991). The absence of cracks also indicates that the remains did not weather in the soil after burial, suggesting that sub-surface weathering was minimal.

10.6. Secondary non-biological modification of vertebrate remains - transport and abrasion

Transport is one of the most important processes which can affect vertebrate remains prior to their stabilisation and final burial. One of the most fundamental aspects of taphonomic research is working out how the faunal components of an assemblage came to be in that place and therefore, the effects of any transport should be determined upon individual remains. Transport of vertebrate material can take place before soft tissue decay, but in the case of most taphonomic studies, only transport of bones is considered. Transport results in the disarticulation and dispersal of bone material away from the life association, carrying remains away from their original environment and in some cases mixing bones from different habitats and even time periods (i.e. time-averaging). The amount and effects of transport depend upon how long the interval was between loss of soft tissues and final burial.

There are essentially two types of transporting agents, physical and biological (Behrensmeyer, 1991). Biological transport includes active disarticulation by predators and scavengers, and passive movement by bioturbation in soils and sediments for buried remains, both of which may cause dispersal and accumulation of bone material (Badgley, 1986). Biological disturbances of bone material and their taphonomic effects have been dealt with earlier (sections 10.3-10.4). Physical processes which may cause scattering and concentration of vertebrate remains include transport by water currents, wave action, wind and gravity. Dispersal patterns produced by fluvial currents have been studied by many authors including Voorhies (1969), Wolff (1973), Dodson (1973) Behrensmeyer (1975), and Korth (1979). Transportation also causes erosion and damage to bone material, which may preserve in fossilised remains. Finally, sorting different body parts according to their hydraulic properties in a waterborne regime produces characteristic bone assemblages which may provide information on their provenance and taphonomic histories (Behrensmeyer, 1991). Microvertebrate remains may be transported by wind action on beaches, dune fields and dried up river beds (Behrensmeyer, 1991). Gravity assists bone movement on steep slopes or at the entrances to caves or sink-holes. The low density of bones means that they can stay at the surface of sediments and thus are exposed to soil creep, mass movement and slope

wash. The Hornsleasow doline may have acted as a depositional low for cadavers and isolated bones. The various effects of transport on the Hornsleasow bone assemblage are considered separately below.

10.6.1. Fragmentation and abrasion

Transportation of bone material away from the skeleton can cause breakages and abrasion to occur. Breakages resulting from transport are indistinguishable from those made by other taphonomic agents, but abrasion damage can be fairly diagnostic. Abrasion of bone material occurs because of the impact of wind or water borne particles on bone or vice versa (Korth, 1979; Shipman, 1981). Therefore both aeolian and hydraulic transport of bones produce pitting and abrasion. In both cases edges of breakages and processes of elements become worn down, rounded and eventually obliterated. Abrasion will obliterate trampling scratches and tooth marks (Shipman & Rose, 1983). In addition severe abrasion can cause removal of the outer surfaces of the bone exposing the inner cancellous tissue to produce 'bone pebbles' (cf. Boaz, 1982; Fiorillo, 1988: Fig. 10.16e-j).

Shipman & Rose (1983) outlined several factors effecting the rate and nature of abrasion damage on bone material, which are:

(1) Grain size and composition of the enclosing sediment. Larger grain size, more durable mineralogy's and increased angularity of sediment particles causes faster abrasion of bone material in experimental conditions (Shipman & Rose, 1983).

(2) Condition of bone (i.e. whether it is fresh, weathered, or mineralised) and the presence of soft tissue or organic matrix on bone. Experiments in tumbling barrels, simulated rotary flume tanks and natural rivers suggest that it takes considerable bone-sediment interaction to cause damage to fresh bones and teeth. Behrensmeyer (1991) showed experimentally that fresh bones could travel over 3km in a sand and gravel river bed without significant damage occurring. Weathered bones more easily abraded (Behrensmeyer, 1991), than fresh material and this is because the slightly elastic surface of fresh bone retaining organic material absorbs some of the shock of the impacting particles (Martill, 1990). Fossilised bones and teeth are less susceptible to abrasion damage, for example Argast *et al.* (1987) tumbled fossilised dinosaur and crocodile teeth for a period equivalent to a distance of 360-480km with coarse sand, and these showed negligible damage to the enamel surfaces. This has a bearing on determining taphonomic histories for time-averaged assemblages containing reworked pre-fossilised material.

(4) The presence of water in the system. Fluvial transport tends to abrade all surfaces of a bone equally (Lyman, 1994), although bone shape can effect this. For instance the edges of bones tend to be worn down more rapidly than their surfaces and convex surfaces (e.g. vertebrae centra) tend to be abraded more thoroughly

than concave or flat ones (Andrews, 1990). Aeolian sand-blasting produces severe etching of the surface of the bone, rather than polishing and rounding (Brain, 1967) and it also tends to abrade only the top surface of an element (Shipman & Rose, 1983).

(5) The duration or distance of transport. Even moderate transport causes microscopic features such as scratches to be quickly erased, but takes up to 35 hours continuous tumbling to induce gross surface stripping (Shipman & Rose, 1983). Although there is a general rule that the longer a bone is transported, the more likely it is to become abraded, this is not necessarily true in all cases. For instance, stationary bones may be 'sand-blasted' by water or wind borne sediment and heavily abraded without significant transport. Also bones can become abraded and polished by excessive trampling on a hard or sandy substrate without significant dispersal, especially if the bones were previously weathered (Brain, 1967). However, experiments have shown that it would take considerably more aelion abrasion to produce any significant rounding of broken edges (Shipman & Rose, 1983).

Several authors (e.g. Shipman, 1981; Fiorillo, 1988) have produced scales of increasing abrasion damage based upon on assessment of angularity or roundness of sedimentary particles which are supposed to reflect to the degree of transport, as observed from tumbling experiments of non-mineralised bones (Table 10.2). Tumbling experiments are notorious as not accurately simulating mechanical abrasion by fluvial or other means. For instance, Shipman & Rose (1983) point out that there is more impacting of particles in a tumbling barrel than in a natural environment and that they have a constant velocity. Tumbling also tends to wear sedimentary particles unequally, producing elliptical, rather than spherical pebbles (E. Cook, pers. comm., 1992). Investigations of fresh, weathered and mineralised bone tissue has been under investigation at Bristol University by E. Cook and suggests that water-borne related abrasion might be much more complex than once thought. Nevertheless these abrasion scales do provide a useful guideline for taphonomic investigations and that of Fiorillo (1988: Table 10.2) was applied to the Hornsleasow bone assemblage. The results are discussed below.

10.6.2. Abrasion stages for the cetiosaur remains

The results of the abrasion investigation for the macrovertebrate remains are shown in Fig. 10.15a and indicate that the cetiosaur skeletal parts show moderate stages of sedimentary rounding (77% in categories '1'-'2', poorly defined peak). These results would suggest that the bones had been transported as isolated sedimentary particles over a reasonable distance (cf. Fiorillo, 1988). For instance, the neural complex figured in Fig. 10.7e shows complete rounding of all processes to form a remnant, and could be called a bone pebble (abrasion index '3'). However, the ilium and vertebra (Fig. 10.7a-d) show only rudimentary damage around their broken edges and upon processes and it

ABRASION INDICES

Based on degrees of sedimentary particle rounding:

'0' = "Very angular". Bone fresh, unabraded.
Processes, edges, and breakages well defined and sharp.

'1' = "Subangular". Slight abrasion. Edges,
processes and breakages show rounding and may show polish.

'2' = "Subrounded". Moderate abrasion.
Processes recognisable only as protusions
on bone. Edges, processes polished.

'3' = "Rounded". Very abraded, bone fragments.
Processes absent or remnant.

Extreme case = "bone pebble" (Plate 1*)

Table 10.2. Abrasion indices for bones based upon degrees of sedimentary particle rounding, after Fiorillo (1988a).

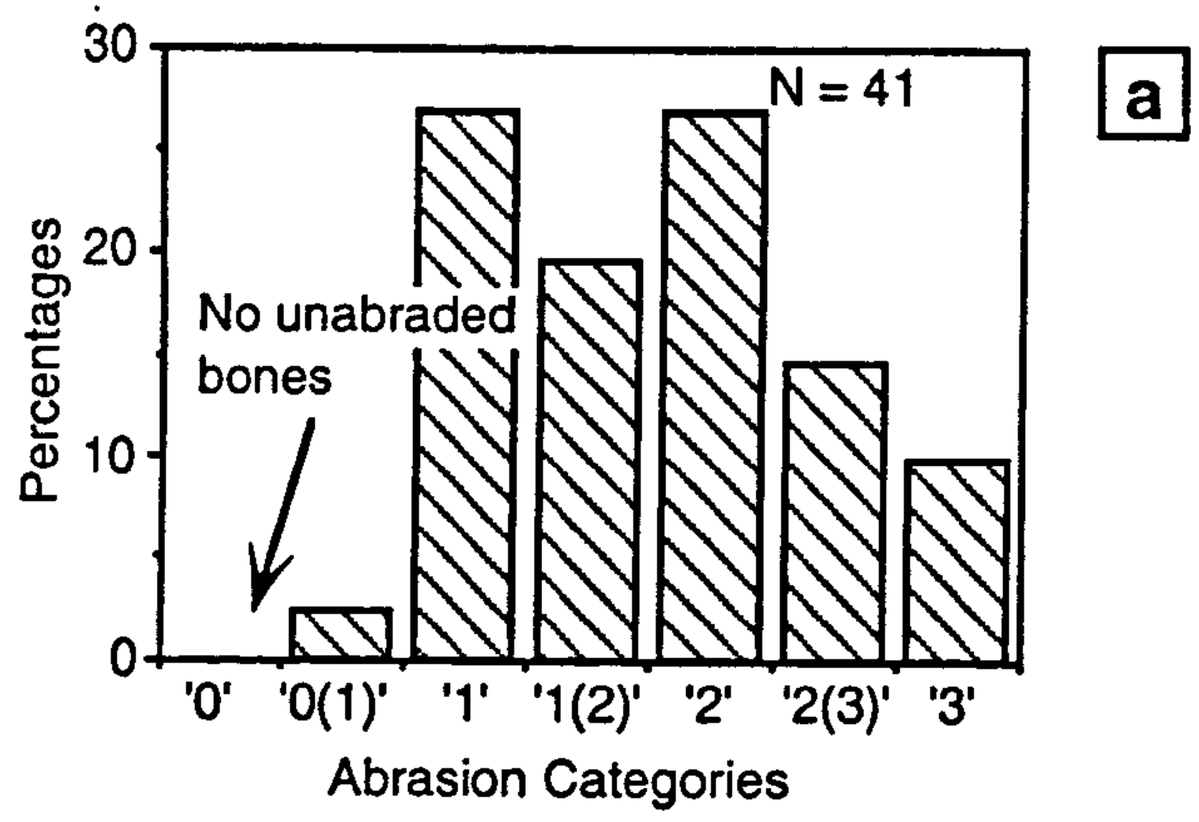
is unlikely that such large and complex parts could have been transported far by non-biological processes (Metcalf, 1993). Evidence from analysis of the orientation and dispersal of the bones, also suggests that there was little fluvial activity in the base of the hollow, and therefore the skeleton although scattered by biological agents appears to be *in situ*.

The apparent discrepancy between the abrasion profile and sedimentary evidence can be explained by examining the surface features of the bones themselves. The graphs showing weathering and abrasion profiles for the macrovertebrate remains are almost identical in form, and most of the abrasion (like weathering) is apparent only upon the upper surfaces. Rounding of the processes on the neural complex has led to exposure of the inner cortex bone and subsequent exfoliation, which in turn has led to crumbling and more rounding of the bone to a pebble. Indeed, the weathering and abrasion of the cetiosaur skeleton seem to be intrically linked, and I believe that much of the apparent rounding has been caused by aelion sand-blasting and karstic erosion in the hollow. The depositional environment for the bones is within a karstic hollow with little soil cover, and it is likely that bone pebbles like the neural complex, would have rolled or been kicked around the base, before becoming mired and stabilised by soil cover and vegetation.

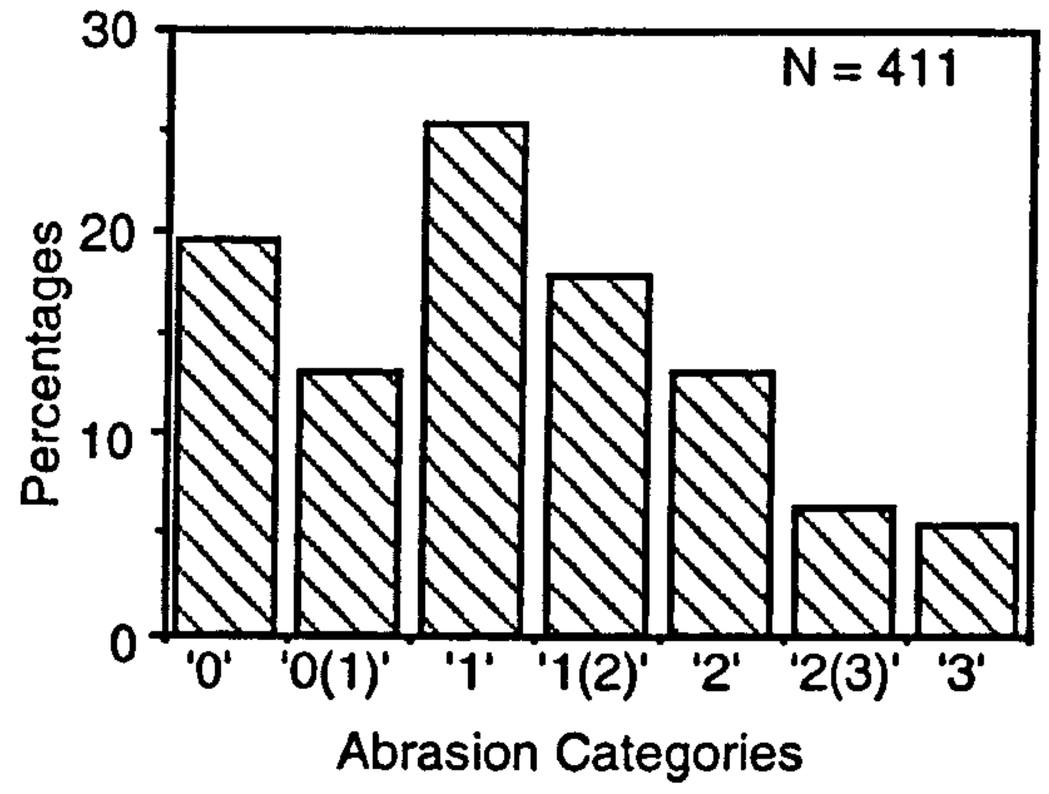
10.6.3. Abrasion profile for the microvertebrate assemblage

The microvertebrate remains show a broad abrasion profile peak, which indicates that in general rounding was moderate (68% in abrasion stages '1'-'2': Fig. 10.15b). These results are similar to those produced by the cetiosaur remains, but this may be coincidental rather than reflecting a common source, as such a broad span suggests a mixed assemblage (Shipman, 1981). The teeth were separated out from the results (Fig. 10.15d) as these more resistant elements tended to be much less abraded than the bone material (76% in abrasion categories of '0'-'1') so skewing the graph. Teeth and especially reptile teeth, are dense and streamlined, and would tend to resist transport and its damaging effects.

The abrasion profile and stages exhibited by some remains (Fig. 10.16) suggests that hydraulic sampling of bones from the surrounding floodplain might have been responsible for bringing some of the material into the hollow. This may have occurred from river or flood-water transportation. However, abrasion of microvertebrate remains may also be controlled by weathering (as outlined above), roll, pedogenic and biological destruction (cf. Maas, 1984). Microscopic abrasion features of the remains were also examined under SEM to attempt to elucidate their cause (Fig. 10.16.). Sliding or forced abrasion tends to removes incompletely mineralised collagen fibre bundles on fresh bone, producing smoothed surfaces with microscopic structure obliterated (Bromage, 1984). Such surfaces are exhibited by tooth wear facets which



a



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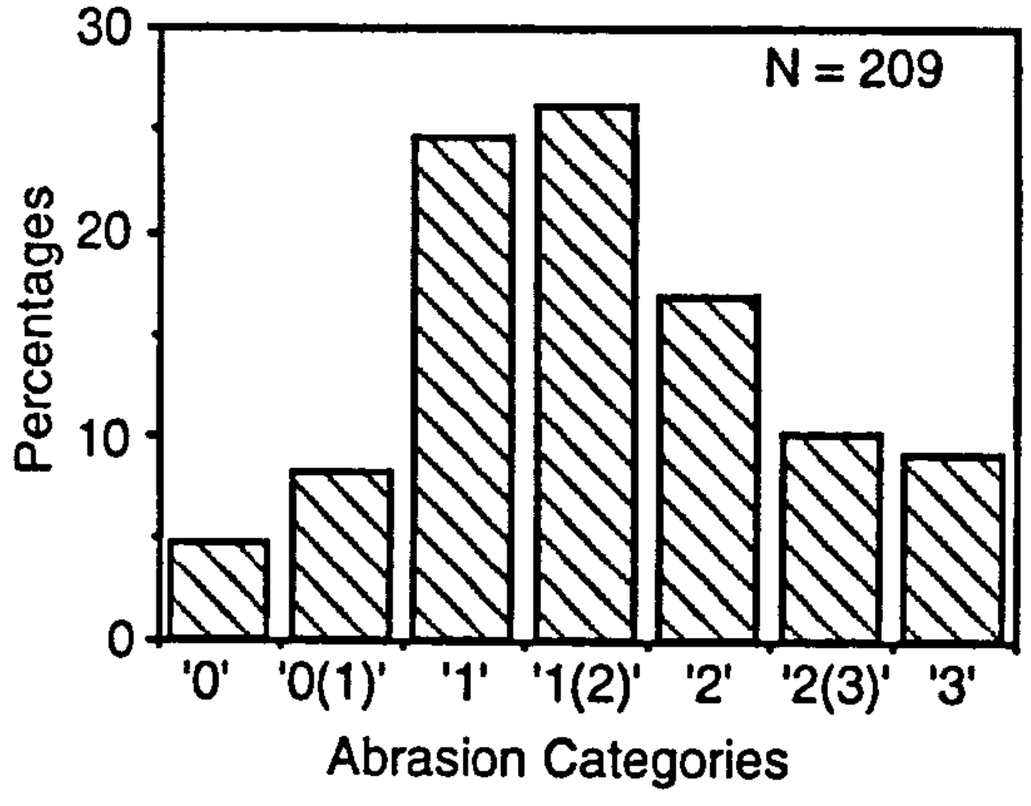

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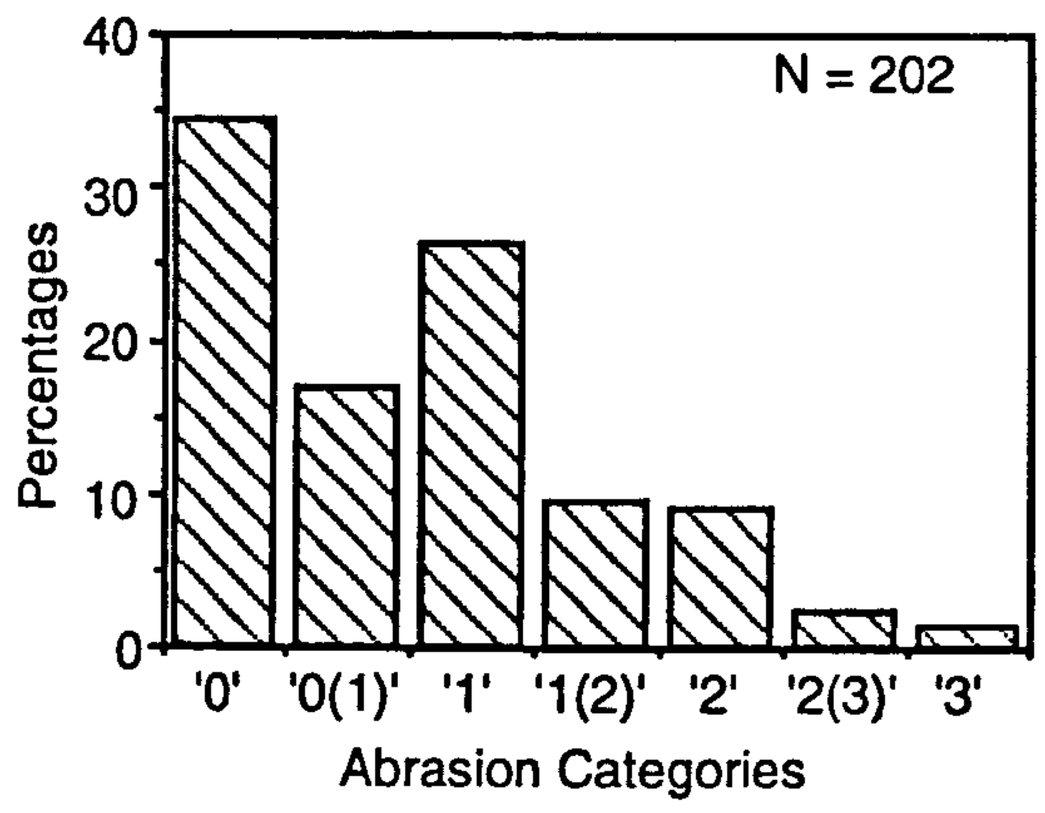


d

Figure 10.15. Physical rounding and abrasion of bones, as exhibited by the abrasion index charts for the macrovertebrate remains (a), the total microvertebrate remains (b), the bones, dermal armour and fish scales in the microvertebrate assemblage (c) and the teeth (d). The number (N) of specimens in each study is indicated.

have undergone progressive forced abrasion and this can still be seen on fossilised teeth (Fig. 10.10g,h). Whereas impacting particles or water jets produce rough abraded surfaces, which appear shiny under the SEM as a result of their increased reflectance from many new surfaces (Fig. 10.16k; Bromage, 1984). Most of the broken edges and processes which have undergone rounding by abrasion show roughened surfaces when examined under the SEM, although they do not appear to be 'shiny' (Fig. 10.16l). Microscopic features are preserved in a number of abraded vertebrate remains, indicating that abrasion was caused by imparting rather than forced wear.

Korth (1979) tumbled micromammalian elements for several hours and saw rounding and eventual breakage due to thinning of the bones by abrasion. Thinning is not observed on any of the Hornsleasow microvertebrate specimens (Fig. 10.16), although this might be because they have disintegrated prior to fossilisation. Very severe abrasion of microvertebrate remains might in fact cause small bones to be destroyed entirely, giving an under-representation of the higher categories in an abrasion profile.

10.6.4. Hydraulic equivalence of bone elements

Unfossilised bones are fairly light, with high surface area to volume ratios and rather irregular shapes. Bone mineral (hydroxyapatite) has a density of about 3.2, whilst individual bones have densities ranging from 1.0 to 1.7, because they contain pore spaces and organic components (Behrensmeyer, 1975). Bones with a specific density of 1.0 will float in water and fresh bone with air-filled pore spaces may remain afloat for hours to days (Behrensmeyer, 1975). This makes them susceptible to transportation by moving water (Behrensmeyer, 1991). Weathered bones which have lost their organic structure, are less buoyant than fresh bones and hence, are less likely to be transported by water (Behrensmeyer, 1991), although they can be moved along the bottom by saltation. Similarly teeth, which have densities approaching 2.0, require stronger currents for waterborne transport than the bones of equivalent sized animals.

Experimental work by Behrensmeyer (1991) has shown that currents of between 20-30cm/s can transport bones from small or medium sized mammals (rodents to sheep), but much stronger currents are needed to move bones of larger animals (e.g. cattle, elephants). She also stated that in natural rivers with a flood velocity of 1.0m/s, bones might be transported up to or over a kilometre distance in a single year (Behrensmeyer, 1991). Bones may continue to move along the bottom of a river system once they have sunk. In a river system the most likely place for final burial of vertebrate material is within the depositional or aggrading parts of the system, such as a point bar or sand bar (Behrensmeyer, 1975).

The hydrodynamic properties of bones in water can be resolved by considering them as sedimentary particles. They can then assigned a "quartz equivalent", which is

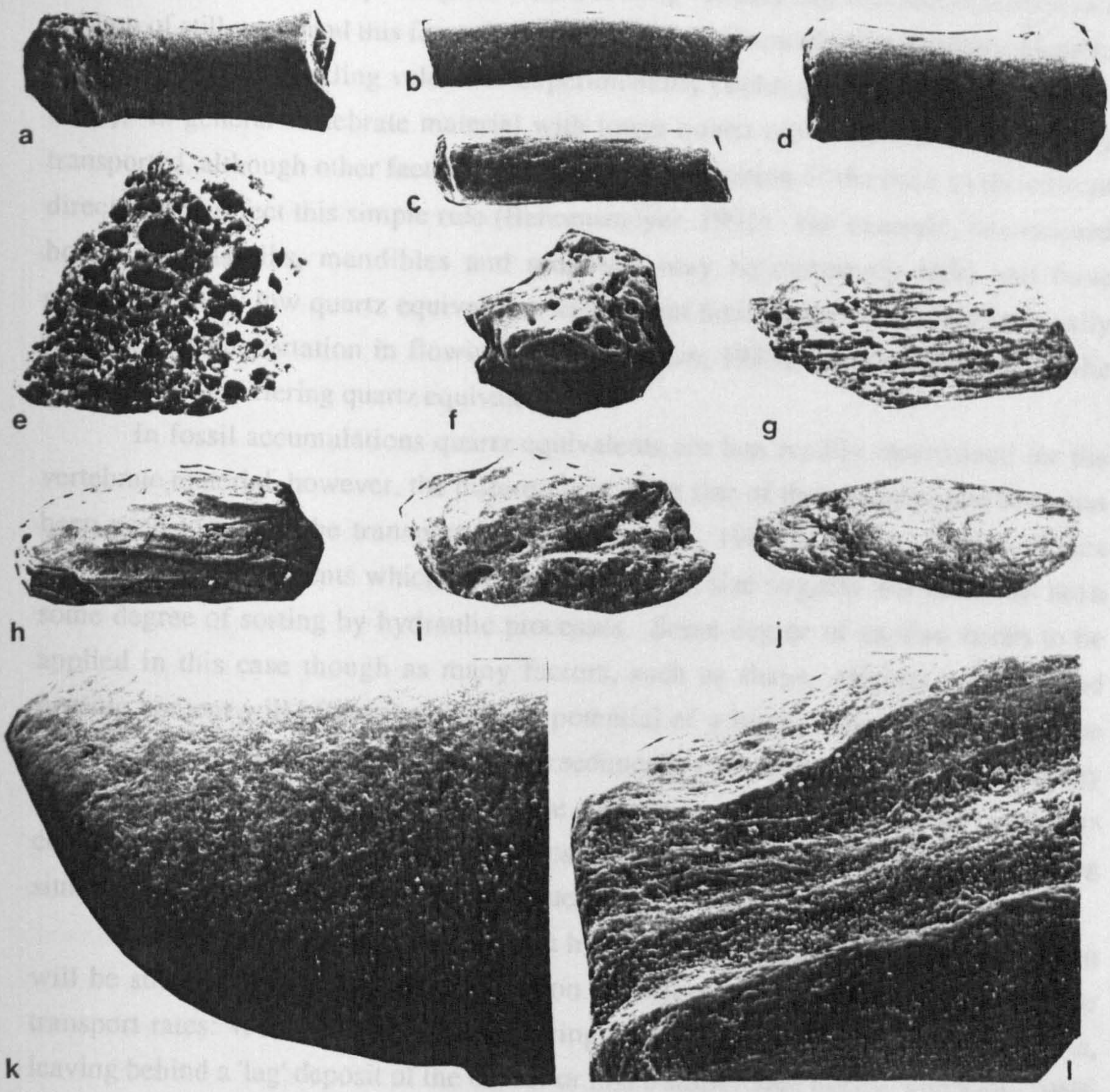


Figure 10.16. Bone abrasion - long bones: (a) G. 52118, broken phalanx, abrasion stage '0' (length = 4.4mm); (b) G. 56908, broken rib, stage '0'('1') (length = 3.7mm); (c) G. 56915, rib fragment, stage '1'('2') (length = 3.2mm); (d) G. 56917, limb fragment, stage '2' (length = 3.4mm). Cancellous tissue: (e) G. 80900, bone fragment, stage '1' (10mm across); (f) G. 77753, bone fragment, stage '1'('2') (2.7mm across); (g) G. 80901, bone pebble, stage '2' (18mm across); (h) G. 78204, bone pebble, stage '2' (4.4mm across); (i) G. 78202, bone pebble, stage '3' (4.3mm across); (j) G. 80902, bone pebble, stage '3' (18mm across). Microscopic details of abrasion: (k) G. 56910, smooth abrasion stage '2'('3') breakage (limb fragment) (field of view = 0.75mm); (l) G. 56911, rough abrasion stage '2' breakage (limb fragment) (field of view = 0.7mm).

defined as the size of the quartz grain with a settling velocity equal to that of a bone in a column of still water and this factor is determined experimentally for each bone element by measuring their settling velocities experimentally (Behrensmeyer, 1975; Shipman, 1981). In general vertebrate material with lower quartz equivalents are more easily transported, although other factors like shape and orientation of the bone to the current direction can effect this simple rule (Behrensmeyer, 1991). For example, streamlined bones such as ribs, mandibles and scapulae, may be extremely light and have correspondingly low quartz equivalence values, but their shape makes them virtually immune to transportation in flowing water (Dodson, 1973). Bones and teeth of the same size have differing quartz equivalents.

In fossil accumulations quartz equivalents are less readily determined for the vertebrate material, however, the difference in grain size of the sediment and bone has been used to assess the transport history (Shipman, 1981; Badgley, 1986). Bones preserved with sediments which have a similar grain size suggest that there has been some degree of sorting by hydraulic processes. Some degree of caution needs to be applied in this case though as many factors, such as shape, original porosity and organic content will effect the transport potential of a bone. Also the nature of the sediments depend as much on the available sedimentary source material as on the energy of the system. At Hornsleasow the presence of so much clay suggests contemporaneous weathering of mudstones and/or ashes rather than the hollow being situated at the distal reaches of a fluvial system.

Because some bones have different hydrodynamic properties, a single skeleton will be subjected to sorting (i.e. separation of different elements according to their transport rates: Wolff, 1973) and winnowing (i.e. removal of the lighter body parts, leaving behind a 'lag' deposit of the denser or less transportable bones: Voorhies, 1969; Korth, 1979; Behrensmeyer, 1991). Therefore distinct patterns of sorting for a fossil deposit can indicate the interaction of currents with a localised bone source. However, input from several sources e.g. several rivers or current and wave interaction along beaches can obscure the patterns of sorting for individual carcasses. Also body parts from many different animals may co-occur in deposits formed by current action (Hanson, 1980) making elucidation of individual skeletal sorting extremely difficult. Such is the case at Hornsleasow were many hundreds of individual skeletons have been disarticulated and mixed in the microvertebrate assemblage. However, a rough approximation of waterborne sorting can be determined and is outlined below.

10.6.5 Evidence for fluvial winnowing - Voorhies Groups and tooth/vertebra ratios

Experiments in a flume tank and in natural rivers by Voorhies (1969), Behrensmeyer (1975) and Shipman (1981) have demonstrated there to be three distinct groups of transport for modern mammalian bones of medium to large sizes. These are known as

'Voorhies Groups' and are illustrated in Table 10.3. Behrensmeyer (1975) discussed dispersal patterns of bone assemblages based upon an analysis of which Voorhies Group bones were present. For instance, the absence of Group 1 bones in an accumulation suggests it has been winnowed and is a fluvial-lag deposit. Furthermore the proportions of bones within different Groups in a fluvially-controlled assemblage might provide evidence for the proximity of fossils to their original death position and habitat (Behrensmeyer, 1975). Korth's (1979) experimental work on the settling velocities of microvertebrate bones and Dodson's (1973) research on their transport in fluvial systems suggests a close similarity of bones into Voorhies Groups, irrespective of their size.

For large accumulations comprising several different taxa, Behrensmeyer simplified this method to examining the number of teeth to vertebrae as a quick index to fluvial winnowing in mammalian assemblages. This method is based upon the fact that in an ungulate mammal skeleton the tooth/vertebra (T/V) ratio is roughly 1.0. However, during fluvial winnowing of the light and high surface area vertebrae from a bone association, a disproportionate amount of the dense, and stream-lined teeth will be left behind in the lag deposit (Badgley, 1986). Thus the ratio basically reflects the energy of the particular portion of a fluvial regime (Fig. 10.17), for example, the channel, levees and point-bars in a river system should show high T/V ratios; whilst at the low energy depositional end of the fluvial system both teeth and vertebrae are deposited in equal proportions, giving a low overall T/V. This has been shown for several modern and Tertiary mammalian accumulations formed by fluvial currents to be correct (e.g. Behrensmeyer, 1975; Behrensmeyer & Dechant-Boaz, 1980; Badgley, 1986; Fiorillo, 1988).

The high proportion of teeth in the Hornsleasow microvertebrate assemblage might suggest that these drag-resistant elements were left behind in the hollow by fluvial winnowing. A living reptile such as the crocodile has approximately 60 teeth at one time and around 70 vertebrae (Romer, 1956), giving a T/V ratio for a single carcass as 0.86. However, as reptiles continually shed and replace their teeth throughout their life spans, they may produce thousands of shed teeth and make this useful rule essentially meaningless. Indeed, at Hornsleasow the T/V of all the teeth (Nt=152; Nv = 3) was 50.83 and even that for the non-shed teeth (with roots attached: Ns = 17) was still extremely high at 5.67, even though the sedimentary evidence suggests a low-energy, essentially floodplain environment (T/V = 1: Fig. 10.17). As jaws are also a component of channel lag deposits (Voorhies Group 1), I have used jaw/vertebra ratios as an alternative. For the champsosaur, *Cteniogenys* (an aquatic and hence, probably indigenous reptile, which possesses distinctive post-cranial and mandibular elements even when they are broken) the J/V ratio in the assemblage (0.10) is roughly equal to that in the skeleton (0.11, based on *Lazarasuchus*, Hecht, 1992: Fig. 10.18).

Group 1: easily removed by low velocity currents; high surface area to volume ratios.	Intermediates	Group 2: gradually removed by moderate currents; low surface area to volume ratios.	Intermediates	Group 3: lag deposit, moved only by high velocity currents; low surface area to volume ratios.
Ribs	Scapula	Femur	Mandibular ramus	Skull
Vertebrae	Phalanges	Tibia		Mandible
Sacrum	Ulna	Humerus		
Sternum		Metapodia		
		Pelvis		
		Radius		

Table 10.3. Voorhies Groups for hydraulic behaviour of bones and their potential of different bone elements for dispersal by water (after Voorhies, 1969; Shipman, 1981).

A) MAMMAL **T/V = 1.00- 1.40****

B)
1) **T/V = < 0.50****
2) **T/V = 3.47***
3) **T/V = 3.12***
4) **T/V = 3.12***
5) **T/V = 3.12***
6) **T/V = 1.21***

C) **T/V = 1.21***

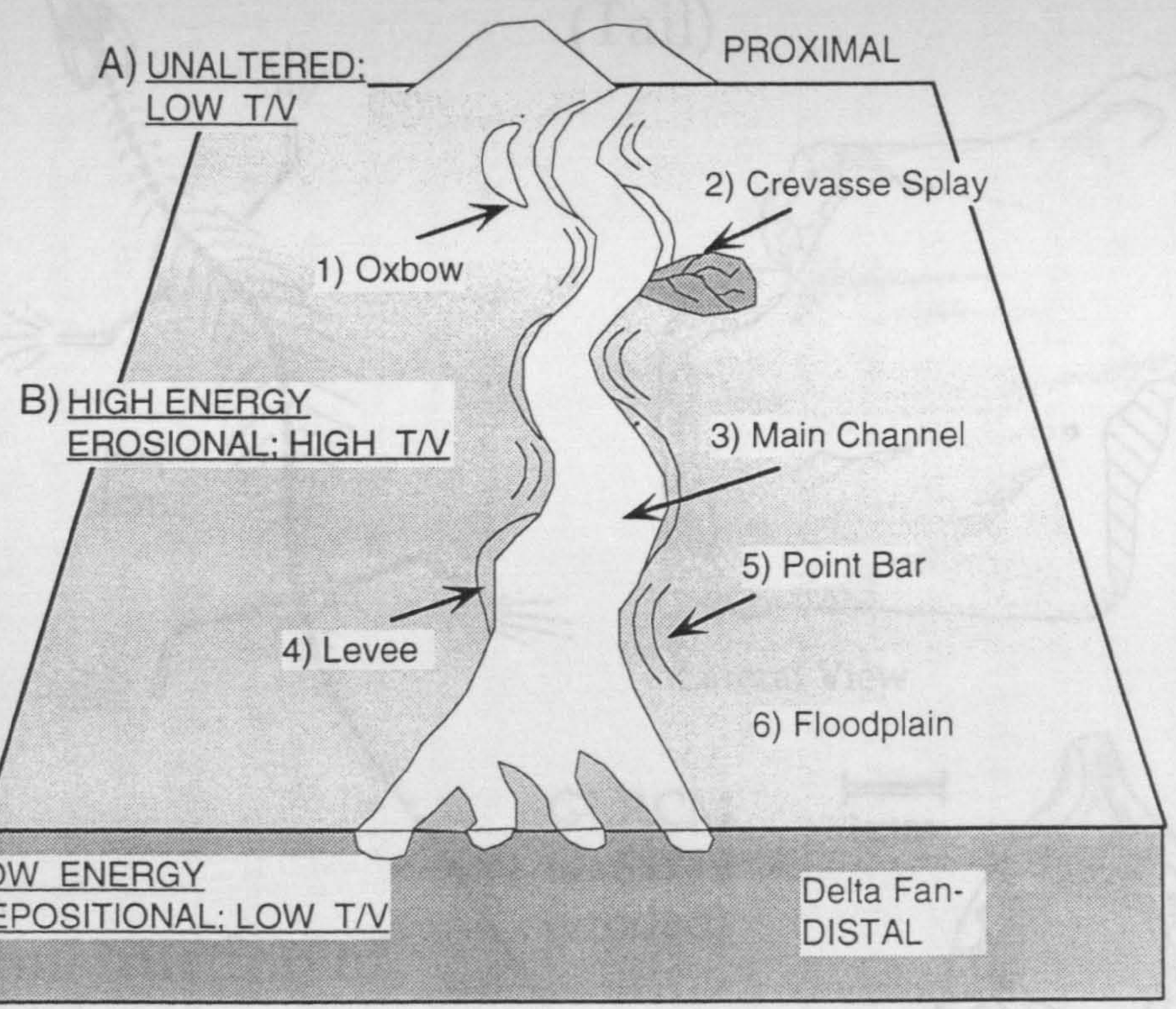
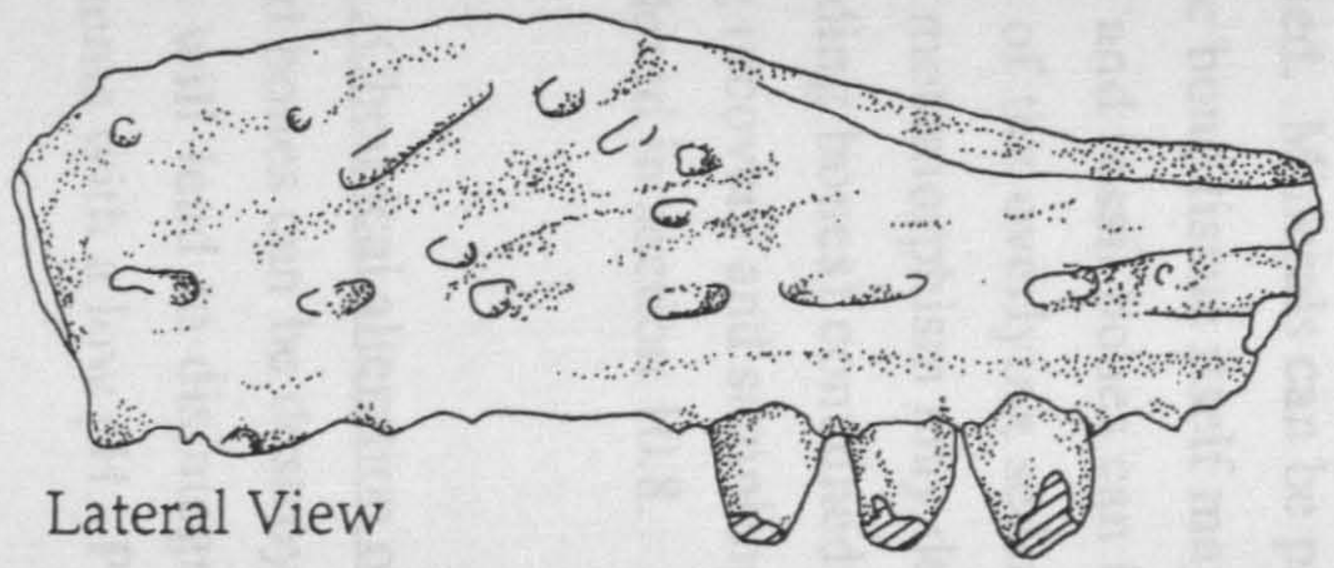


Figure 10.17. An idealised fluvial facies diagram, showing the tooth/vertebra (T/V) ratios in various parts of the river system. The T/V ratios assess whether or not a bone assemblage has been selectively sorted by fluvial currents prior to burial (after Behrensmeyer, 1975; Behrensmeyer & Dechant-Boaz, 1980).

MAXILLAS
(Upper jaws - with teeth)



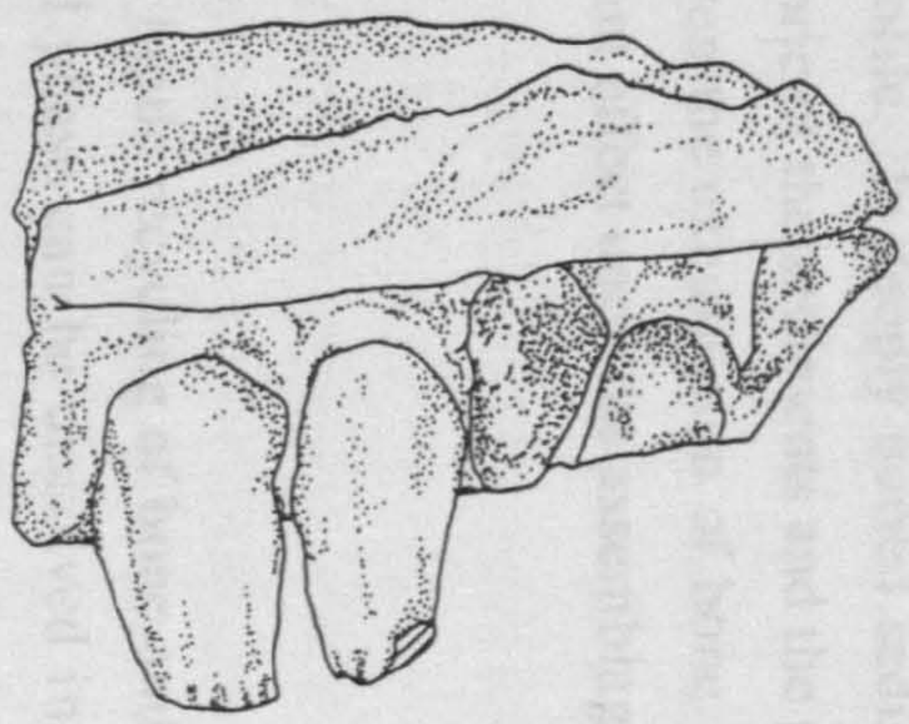
Lateral View

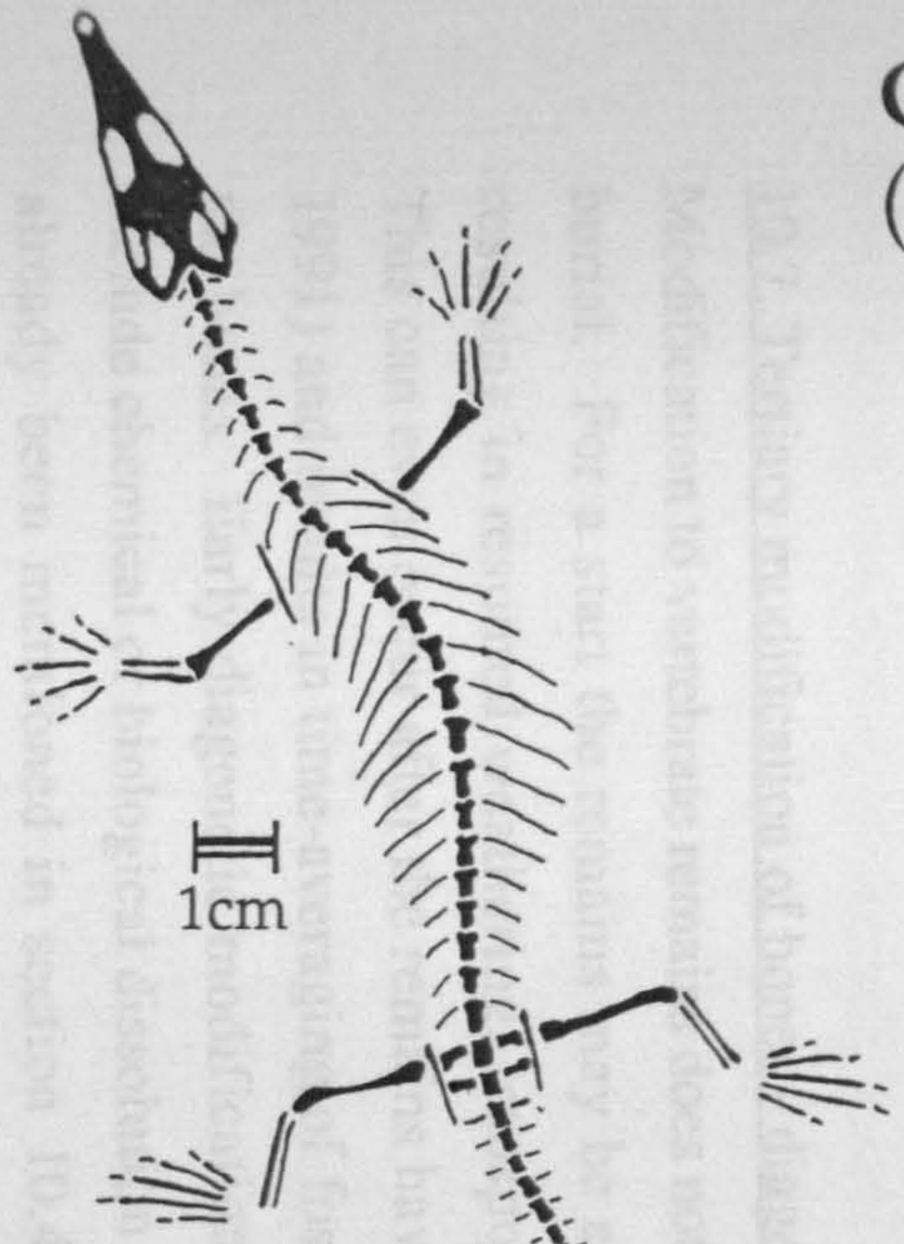
1mm

GLRCM G. 66855
(juvenile)

Medial View

GLRCM G. 66854

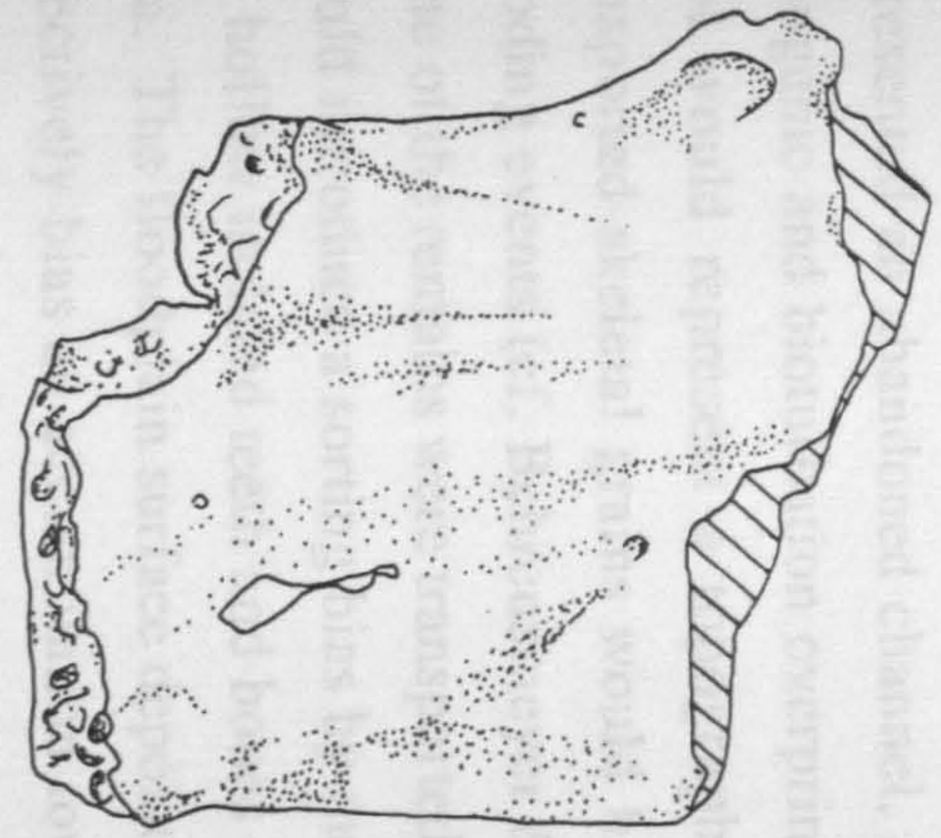




1cm

Reconstruction of
Cteniogenys
(after Evans, 1991)

CAUDAL VERTEBRA
(Tail)

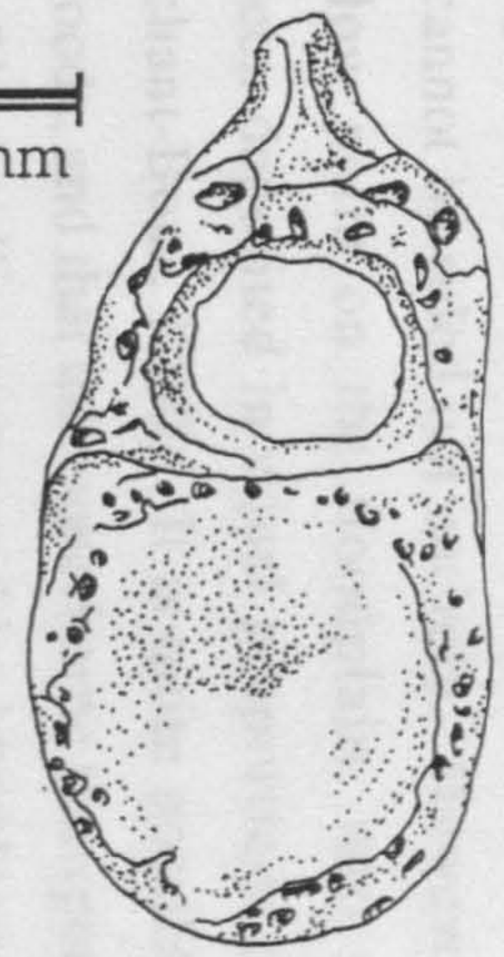


Lateral View

GLRCM
G. 51749
(eroded)

1mm

Front View



JAW : VERTEBRA RATIO

(J/V) = 1 : 10

Figure 10.18. Jaw/vertebra (J/V) ratio for the champsosaur *Cteniogenys*.

10.6.6. Discussion

There is no evidence for fluvial activity in the hollow (section 10.1.2) or that the pool represented an abandoned channel, although destruction of sedimentary features by pedogenic and bioturbation overprinting cannot be ruled out entirely. However, the pool would represent a topographical low point on the floodplain, into which transported skeletal grains would have been deposited in equal proportions during flooding events (cf. Behrensmeyer & Dechant-Boaz, 1980). It is quite possible that some of the remains were transported by floods, and that this waterborne transportation would introduce a sorting bias by means of hydraulic sorting and by introducing into the hollow isolated teeth and bones as well as animal cadavers from the surrounding area. The floodplain surface deposits would have undergone weathering, and this can effectively bias an accumulation toward the more durable skeletal elements such as teeth.

10.7. Tertiary modification of bones - diagenetic and post-burial changes

Modification to vertebrate remains does not stop once the material has been stabilised by burial. For a start the remains may be exhumed and reworked, before final burial, resulting in resumed weathering, transport or other pre-burial taphonomic changes. This can even occur after the remains have been mineralised ('pre-fossilised': Martill, 1991) and results in time-averaging of fossil faunas as well as the physical changes to the bones. Early diagenetic modifications occurring within the enclosing sediments include chemical or biological dissolution of the bone material (root acid alteration has already been mentioned in section 10.4) and movement by bioturbation or other processes. Alteration can occur once the surrounding matrix has been diagenetically modified. Minerals can be precipitated in the voids left by decaying organic material, and the bone tissue itself may be replaced wholesale by a new mineral in some cases. Bones and fossil bones can be deformed during this process and by the compressive forces of the overlying sediments in a buried profile. In deeply buried sedimentary rocks, metamorphism may destroy the original fabric of the sediments and the particles (including bones) contained therein. Finally, even the exhumation of bone material during recovery and sampling can have a significant effect upon the assemblage, this is considered in section 10.8.

10.7.1. Chemical alteration of bones in sediments

Buried bones can be destroyed soon after burial, biotic recycling of bone and organic tissue will lead to disintegration or the mineral phase can be dissolved in soils or sediments with a low pH. Properties of the enclosing sediment, such as its chemistry,

acidity and moisture levels, pedogenic effects and permeability will all effect the rates of alteration. Temperature and moisture levels may govern rates of alteration, especially if these are biologically initiated; their effects are essentially sub-surface bone weathering and are early diagenetic or 'syndiagenetic' processes (Rolfe & Brett, 1969). The properties of the remains themselves, such as size, chemical and structural composition, will also play a part in determining the amount of sub-surface alteration which takes place. For instance, damaged bone material and small bone fragments are much more susceptible to chemical leaching than whole skeletal elements (Klein & Cruz-Uribe, 1984), and hard, compact structures, such as tooth enamel are also more resistant to alteration in the sediment.

Most post-burial chemical alteration of bone material is usually attributed to an acidic microenvironment surrounding the remains, due to organic acids in soils or an acidic groundwater regime (Williams, 1987). For instance, in acidic bogs, soft tissue may be preferentially preserved by the sterilising effect of the low pH conditions, with bones being completely dissolved away. The classic example of this are the 'bog people' human remains from prehistoric peat deposits (Glob, 1969). Complete dissolution of bone apatite by acidic pore-waters could be a relatively common occurrence, but only when soft tissue outlines or when consolidated rock preserves the external moulds of these fossils can dissolution be demonstrated (Martill, 1991). The lack of vertebrate fossils in the Middle Jurassic alluvial and fluvial sediments of the Cleveland Basin could be because of such conditions, as plant material is exceptionally well preserved, whilst invertebrate fossils, such as fresh-water clams are only ever found as moulds in lacustrine limestones. However, no moulds of vertebrate remains have been found, but this may be due to a sampling failure or the unconsolidated nature of much of the floodplain muds. At other fossiliferous localities, such as the lithified Middle Triassic alluvial sandstones and conglomerates of Devon, skeletal material has been reconstructed from latex casts taken from the well-preserved moulds found in the rocks (P. Spencer, pers. comm. 1991). Under favourable chemical conditions, other minerals such as calcite, iron oxides and silica may fill the moulds left behind by dissolved bones.

Apart from the biological acid attack upon buried microvertebrate remains in the Hornsleasow soil, there is no evidence that the vertebrate material suffered any substantial chemical dissolution. The unconsolidated nature of the clays and their processing by wet sieving would destroy void spaces left by dissolved elements. However, although the presence of substantial unoxidised plant matter and a limited amount of pyrite, suggests anoxic microenvironments, where pH may have been sufficiently low to etch bones, the abundance of abundant calcium carbonate in the sediment and the low permeability of the clays would buffered the build up of organic

acids in the Hornsleasow soil and near neutral to alkaline conditions seem to have prevailed (Fig. 7.2).

10.7.2. Compaction and other physical modification

The physical modifications which can alter vertebrate material after burial include the effects of bioturbation and pedoturbation in developing soils and crushing and breakage of bones by compaction due to the weight of overburden upon buried sedimentary profiles. The former is probably only sufficient in affecting small remains, although large plant roots can apply a considerable force to obstructions. Effects such as bioturbation and soil forming processes can lead to physical movement of vertebrate remains, usually in a vertical plane, which may result in a limited amount of reworking or time-averaging. Also articulated or associated material may be separated by soil animals in a well-bioturbated sediment. These are likely to be early diagenetic changes.

Compaction by overlying sediments compresses sedimentary particles (including fossils) in a buried deposit and lessens the pore spaces between them. It is a late diagenetic or 'anadiagenetic' effect occurring during deep burial (Rolfe & Brett, 1969). The compaction of vertebrate tissue depends upon the composition of the material, whether it has been mineralised, the thickness of cortical bone and the bones' orientation in the sediment (Martill, 1991). Martill (1991) reports that in the Lower Oxford Clay vertebrae of marine reptiles may show differential crushing damage due to their complex structure and its response to the downward force. Non-mineralised bone tissue which has been less damaged before burial is much more likely to survive compaction after burial (Klein & Cruz-Uribe, 1984). Resistance to compaction is also greatly increased by infilling of pore-spaces by early diagenetic minerals and mineralisation of the bone tissue itself. Finally, even the composition of the surrounding sediments may limit squashing of bone material; coarse grained sediments tend to be more resistant to compaction than muds and silts (Briggs, 1990).

Shipman (1981) detailed two responses to compressive forces exhibited by bone and fossil bone material. She suggests that rapid compression results in crushing and breakage of bones, whilst slow compression may result in plastic deformation.

Compaction breakage of bone tissue results in the flat, smooth breakage surfaces perpendicular to the long axis and microfabric of the bone, known as 'transverse fractures' (Fig. 10.6f). True transverse fractures are thought to occur after mineralisation (Shipman, 1981) and follow lines of weakness in the enclosing rock or sediment. However, factors controlling the fragmentation pattern of vertebrate remains which have undergone compaction are the composition, structure, shape and size of the bone material itself. Crushing of non-mineralised bones in a sediment would result in the stepped or columnar fracture patterns developing along desiccation cracks and rectangular fragments will be seen in an assemblage (Figs. 10.6h and 10.19a). It is

also conceivable that mineralised bones which had previously been damaged, by earlier taphonomic modifications such as weathering or spiral fracturing could break along already developed lines of weakness in the bone, as well as or rather than along transverse fractures. Therefore, post-mineralisation crushing damage might be difficult to distinguish from that modifying dried bone material (Shipman, 1981), such as bones crushed by trampling on a hard substrate. Cortical bone tissue tends to break cleanly, but even mineralised cancellous tissue will break into irregular splintered fragments (pers. obs.). Tooth enamel tends to break with a conchoidal fracture (Fig. 10.19b), but this sort of chipping may occur at any time during the taphonomic history of the tooth.

However, there are a few pointers which can lead an investigator to deciding upon compaction as the taphonomic agent involved in bone breakage. Bones broken by compaction may show conjoining fragments lying close to one another in the enclosing sediment, which might be re-joined following recovery. Such a task was undertaken upon the badly compaction fractured (and explosion damaged) femur of the *Cetiosaurus* dinosaur (Fig. 10.19c). The smaller the fragments tend to be however, the less likely a bone will be recognised and re-assembled. Also, bulk processing techniques such as the sieving procedures adopted in the Hornsleasow investigation tend to limit the likelihood of recovering associated fragments; although some were found in the residues (for example, the two 'halves' of a fragmentary teleosaur crocodile tooth, split down the centre of the tooth by a large transverse fracture). Other evidence that fractures have occurred after fossilisation (and hence by compaction) is that their breakage surface exposes a lighter coloured (?less mineralised: Shipman, 1981) tissue compared to the outer bone surface. The composition of the skeletal assemblage can indicate preferential destruction by post-depositional forces. For instance, Klein & Cruz-Uribe (1984) have suggested that an abundance of isolated teeth, tooth-bearing elements and dense bones such as phalanges and hand and foot bones suggests post-depositional destruction of other bones. The Hornsleasow microvertebrate assemblage shows such a preferential preservation of some of these elements (section 10.1.3). However, other pre-burial taphonomic agents may result in a similarly structured skeletal assemblage (e.g. fluvial winnowing and predator sorting).

Plastic deformation occurs when the original shape and dimensions of the fossil are changed by slow distortion without breakage. Finks (1979) indicates that it is made possible by micro-fracturing and recrystallisation of the bone material. Some deformation may take place in most compacted bone tissue before final fracturing. Some elements are more prone to plastic distortion, for instance skulls are susceptible to deformation and squashing, because of them being a large, hollow structure (Lyman, 1994).

The cetiosaur bones have suffered a small amount of compaction breakage within the deposit, but no obvious distortion, except along small-scale fault-lines, with

an average displacement of between 1-2mm, but up to 10mm in some specimens. Transverse breakages are common. Some of the bones show considerable squashing damage of there interior cortex. Petrographic thin sections of some of the vertebrae reveal that the bone tissue is 'crazed' by hairline fractures within the vascularised tissue (Fig. 10.19d). In some cases, the tissue has been pulverised to a powdery substance, similar to the fault gouge seen along rock thrust planes and when the outer cortical bone is removed or the bones sectioned, the interior crumbles. Large scale fractures may be infilled and possibly enlarged by calcite veining (Fig. 10.19d).

The difficulties in differentiating between pre-fossilisation and post-mineralisation fractures in microvertebrate remains has been highlighted in section 10.2.2. Regular-sided roughly cubic fragments of bone are formed from breakage of cortical bone (Fig. 10.19e) , but bone splinters are formed from the compression of non-linear elements and tooth dentine. Limb bones tend to break along the grain of the fabric as well as perpendicular to it (i.e. stepped fractures). Tooth enamel is criss-crossed by a network of fractures, some which follow pre-existing longitudinal cracks and others perpendicular to them (Fig. 10.19a). This leads to post-mineralisation exfoliation producing characteristic rectangular flakes of enamel to be lost (Fig. 10.19a). No plastic distortion or shear displacement was seen in any of the studied microvertebrate remains. However, on such small material, shearing resulting from compaction could completely separate the two sides of the fracture. The resulting fragments should be conjoined or in close proximity, but later bioturbation, pedogenic or sampling disturbance can mask this effect.

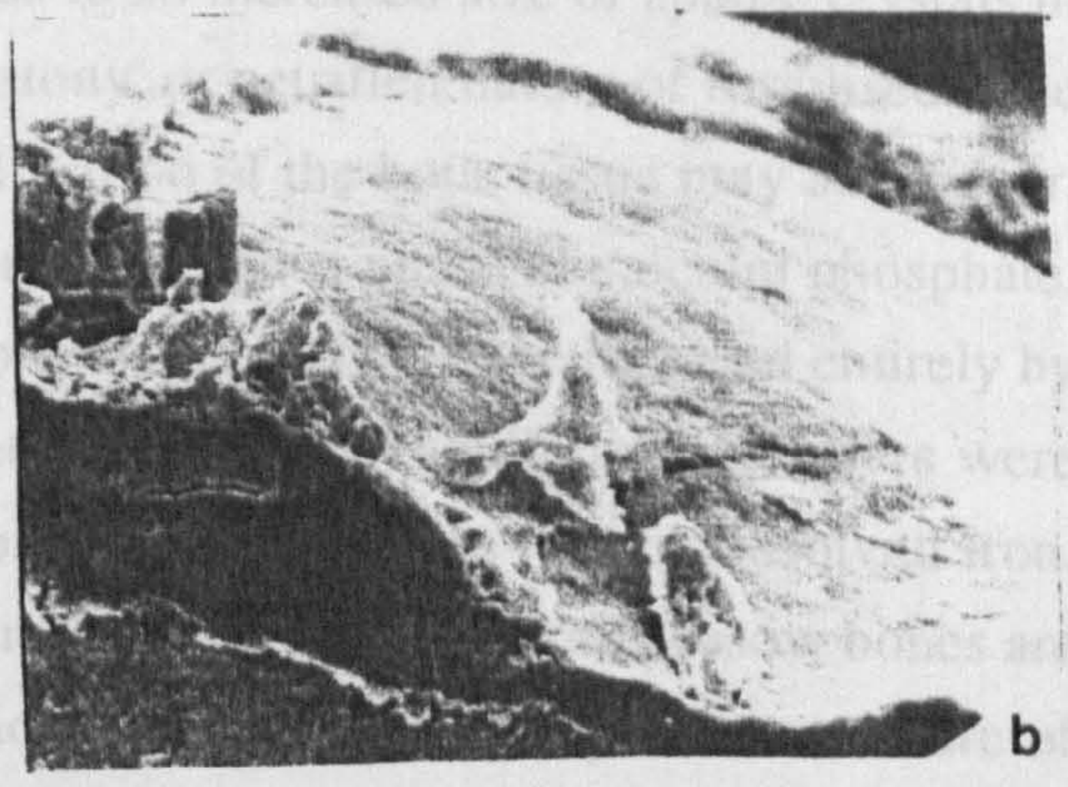
10.7.3. Mineralisation and diagenesis

The diagenesis of vertebrate remains can be exceedingly complex, but essentially takes place as the result of three processes, which were outlined by Martill (1991) as: the diagenesis or mineralisation of the bone tissue itself; permineralisation within the pore spaces and cavities in the bones; and the diagenesis of the surrounding sediments. Diagenesis of skeletal material is affected by the features of the specimen itself, such as its size, porosity, chemical and molecular structure and by the external factors of the enclosing sediment, such as ground water pH, moisture levels, and temperature (Lyman, 1994). Mineralisation can be syndiagenetic in shallow sediments involving biotic chemical reactions, or can occur during later burial, when inorganic chemical reactions predominate (Rolfe & Brett, 1969). Mineralisation of fossil vertebrate material gives it new physical and chemical properties, which might make it more or less susceptible to later taphonomic effects.

When the original mineral matrix of the bone is replaced by another mineral, the process is known as bone 'mineralisation' or fossilisation. In most cases this is simply the ionic substitution of OH with F, or to a lesser degree Cl, and PO_4 with CO_3 in the

Figure 10.19. Post-mineralisation modifications. Fracturing (a-e), in detail: (a) G. 51223, transverse cracking and splitting along pre-existing weathering cracks on the enamel of a tritylodont tooth cusp (field of view = 0.7mm); (b) G. 61601, close up of conchoidal chip at base of crocodile tooth (field of view = 1.4mm); (c) G. 10020, shattered *Cetiosaurus* femur (length = 1.5m); (d) RV.91 (unnumbered specimen), petrographic thin section through *Cetiosaurus* vertebra process fragment, showing calcite vein (1mm diameter); (e) G.61604, close up of preparation damage on crocodile tooth (field of view = 75μm). Microscopic preservation and mineralisation (f-i), in detail: (f) G. 61601, microscopic structure of dentine in broken crocodile tooth (field of view = 40μm); (g) G. 61407, microscopic structure of dentine in crocodile tooth (field of view = 180μm); (h) RV.91, petrographic thin section showing calcite infilled pores in cancellous bone (field of view = 2mm); (i) G. 10015, amorphous pyrolusite crust on cetiosaur bone fragment (field of view = 95μm).

a



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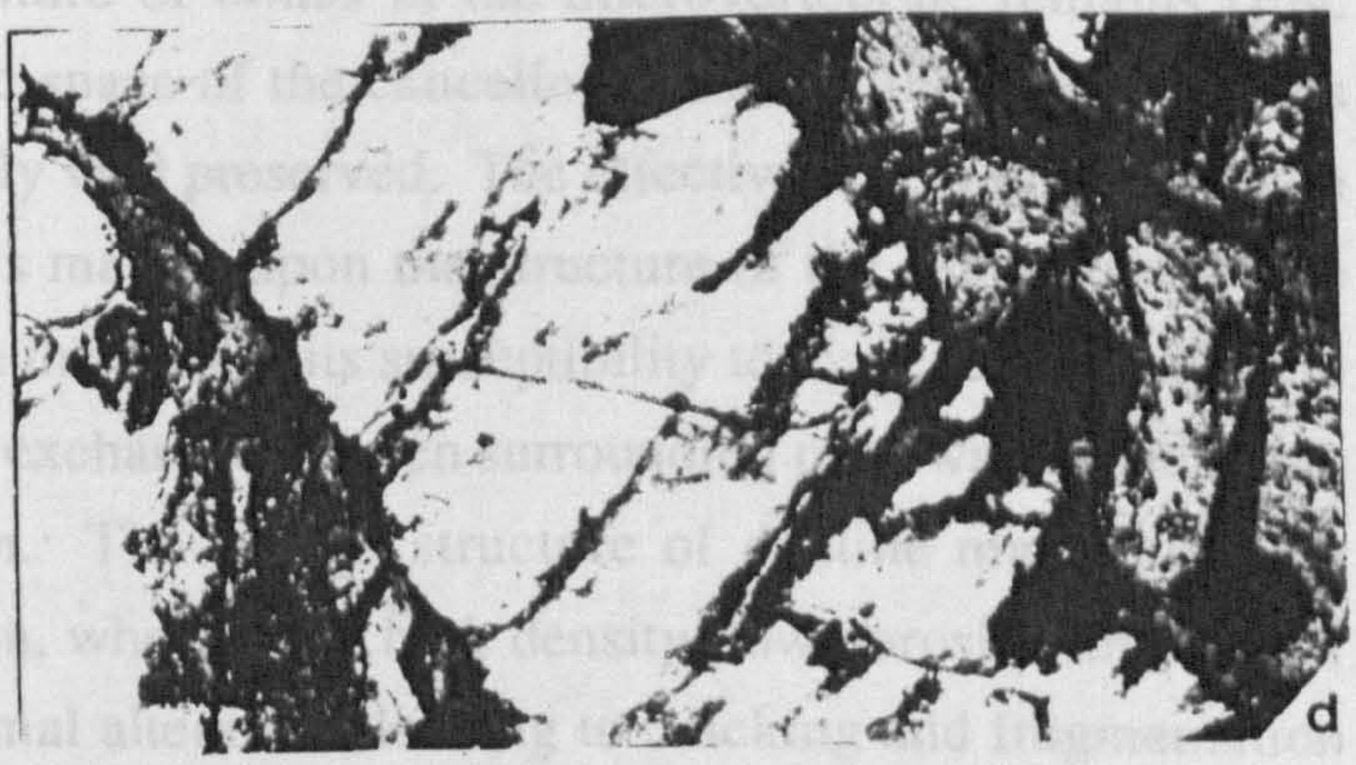

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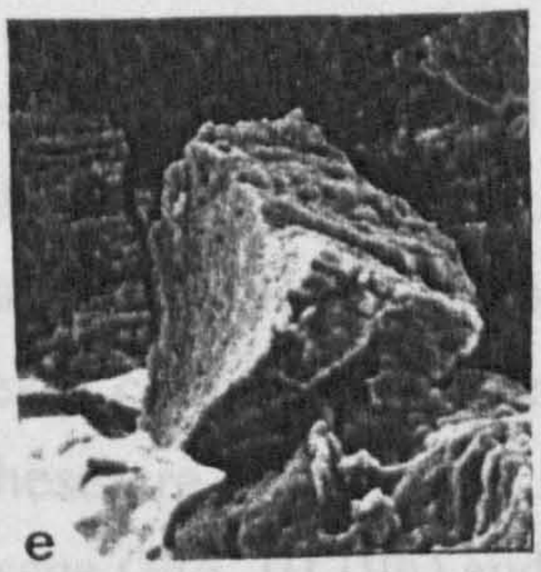

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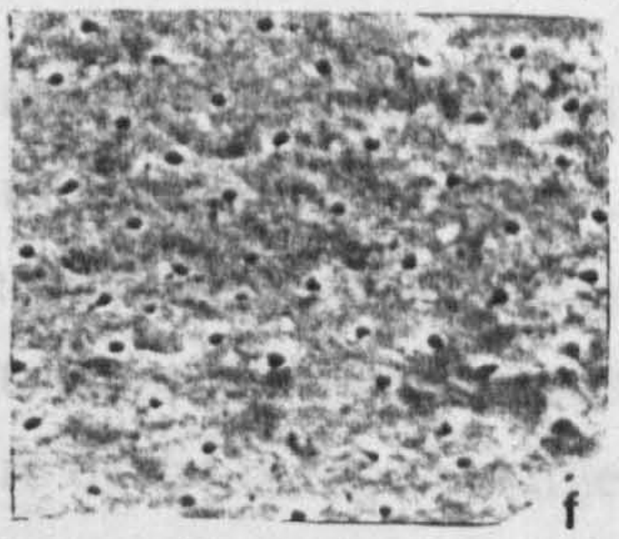

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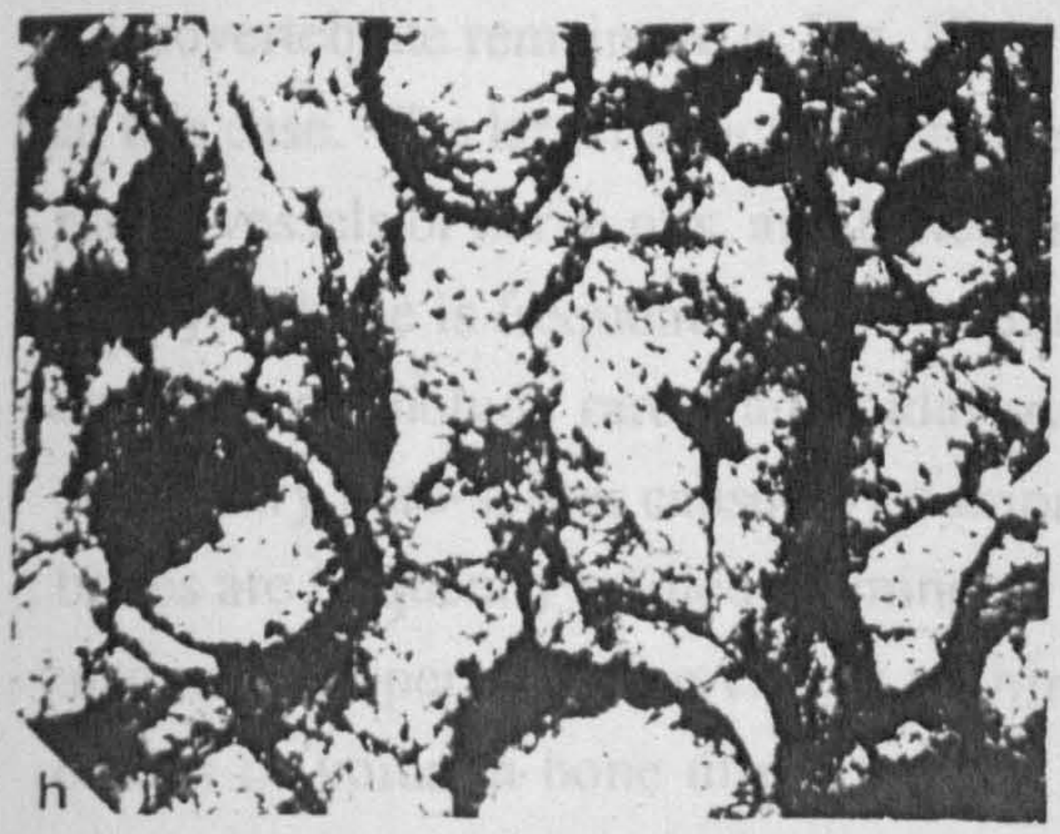

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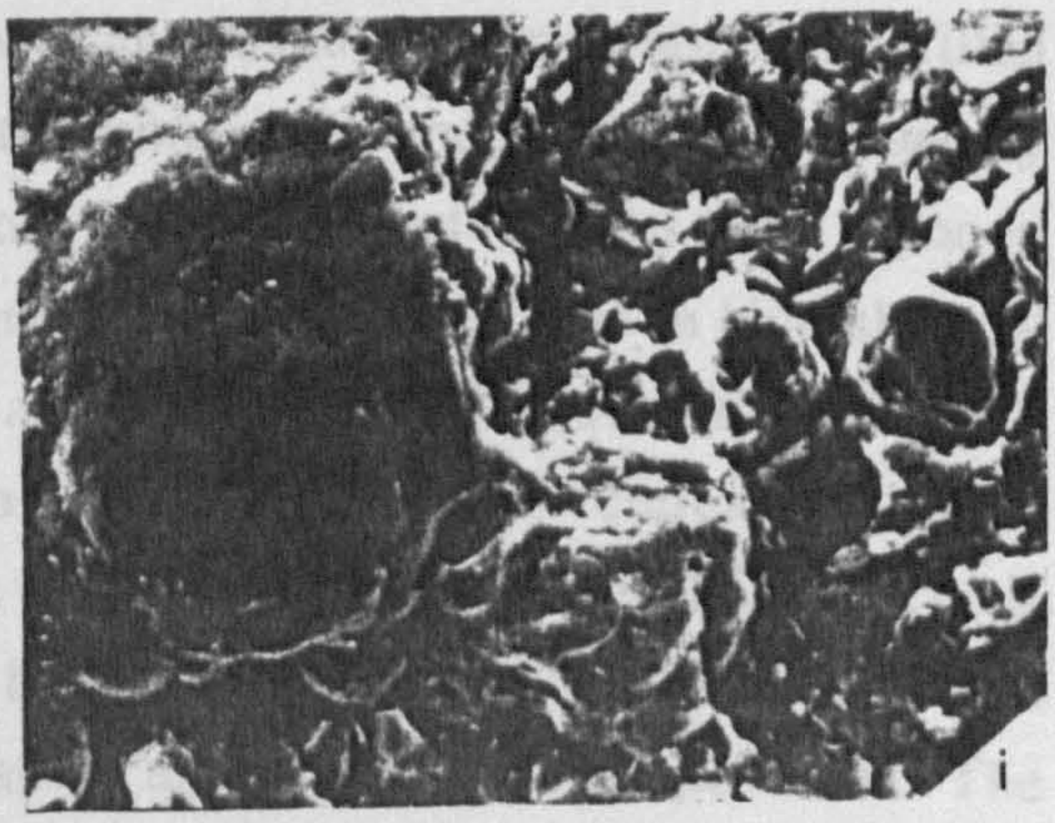


i

bone apatite. This sort of mineralisation leads to an increased size of apatite crystals in mineralised bone, leading to the brittle and 'stony' or petrified nature of fossilised bone (Bartisiokas & Middleton, 1992). Recrystallisation of the bone tissue may also occur, as can an early diagenetic overgrowth with an amorphous phase of calcium phosphate, but both of these are rare (Martill, 1991). Bone apatite may also be replaced entirely by another mineral, for example early diagenetic pyrite, or in acidic groundwaters were phosphorous is liberated from dissolving apatite and which are rich in dissolved iron, the bright blue iron phosphate, vivianite (Martill, 1991). All the Hornsleasow bones are preserved as a hydroxyapatite. Microscopic features such as the porous structure of tooth dentine and the fibrous nature of bones in the microvertebrate remains (Fig. 10.19f,g) and the canals and pore-space of the cancellous tissue in the larger remains (Fig. 10.19h) are also are extremely well preserved. The effectiveness of mineralisation at preserving bone tissue depends mainly upon the structure of the tissue itself. The greater the porosity of bone tissue the greater its susceptibility to diagenetic change and the more rapid the chemical ionic exchange between surrounding pore-waters and tissue can happen causing fossilisation. The porous structure of dentine makes it fairly susceptible to diagenetic alteration, whereas the high density, low porosity structure of enamel means it undergoes minimal alteration, leading to cracking and fragmentation between the two phases.

Minerals precipitated in the voids left in skeletal material following the decay of organic matter can preserve the detailed microscopic structure of the bones or teeth. This process is known as mineral impregnation or 'permineralisation' and provides extra strength to porous bone to resist compaction. There are two spaces which can be impregnated during diagenesis, these are the original structure of the bone, the large pore spaces between cancellous tissue (Fig. 10.19h) and marrow cavities in long bones, and the tiny vessels within the bone itself. The tiny vessels, lacunae and canaliculi are numerous but constitute a smaller volume than the larger cavities and are not well connected to the surface environment except where breakages have occurred and internal structure exposed. Hence they are not often filled and from SEM studies of the microvertebrate remains (e.g. Fig. 10.19f,g), we can see that they have not been infilled in this case. The larger structures are frequently connected to the surface of the bone by blood vessels or nerve pits, and hence, are more usually infilled (Fig. 10.19h). Spongy cavity of bone is frequently filled with diagenetic minerals, and those in Hornsleasow assemblage include rare framboidal pyrite, and two phases of calcite (Fig. 10.19h). Secondary pore-space, caused by compaction or pre-fossilisation fractures in the large bones are frequently infilled by mineral veins (Fig. 10.19d). A late diagenetic calcite is also seen to pervade fine cracks within the bones. Matrix or minerals infilling fine cracks or voids in bone might expand and splay apart pieces of the specimen. The microvertebrate material is not so heavily permineralised as the cetiosaur bones. Within

much of the vascularised tissue and bone cavities there is limonite, and a ferroan calcite cement. Pyrite is not found, but poorly crystalline limonite within the interior of the bones, may be a replacement product of pyrite.

Comparison of infilling mineral suites with those in the surrounding matrix may indicate transport to the final burial place if the two sets are dissimilar (Lyman, 1994). Vertebrate remains which have been mineralised and then reworked to a later deposit are known as 'pre-fossilised' specimens (Martill, 1991).

Diagenesis of host sediment only affects the outer surface of bone. Soluble salts e.g. manganese and iron oxides, and calcium carbonate are derived and transported by ground water (Lyman, 1994), to be precipitated as crusts ('encrustation': Lyman, 1994). A very early diagenetic surface crust of pyrolusite (MnO_2) was precipitated on the cetiosaur remains. The crust is generally amorphous (Fig. 10.19i), but occasionally forms a fine dendritic pattern over the bones and limestone clasts at the base of the hollow. The crust is thought to have been precipitated during the transition from a karstic hollow to marshy soil following deposition of the Grey Clay layer. The crust overlies trample scratches. Other encrusting minerals are limonite and a non-ferroan calcite which is found only on the lower faces of the bones, cementing them to the underlying palaeokarstic surface. Calcium carbonate coatings are thought to form upon lower or downward-facing surfaces first (Lyman, 1994). The microvertebrate remains show limited encrustation by limonite, pyrolusite crusts are not present on the bones, suggesting that they were quickly buried.

10.8. Tertiary modification of bones - sampling damage

Generally much of the vertebrate material sampled and sorted by the excavators, museum, academic and technical staff employed during the 5 year span of the investigation was not modified by their care. However, sampling can introduce certain biases and alterations to an accumulation (for example, fragile specimens may be broken and less attractive or unidentifiable bone slivers might be unrecognised or overlooked during sorting procedures), and although every effort was made in this study to retain and examine all bone, whether identifiable or not, some problems did occur.

10.8.1. Excavation and conservation damage

During the excavation of the clay lens controlled blasting was implemented to remove the overlying limestone. At that time, priority was to remove the clay as quickly as possible to expose the cetiosaur skeleton. By misfortune, one of the explosive charges was placed in the limestone directly above the intact 1.5m long cetiosaur femur and the resulting shock waves shattered the femur into hundreds of pieces (Fig. 10.19c). Through the painstaking care of museum worker, R. Vaughan, much of this damage was restored by careful labelling of each fragment and removal to the laboratory for

preparation. The femur has now largely been re-assembled, and forms part of the permanent exhibition at Gloucester City Museum. Less than 5% of the total macrovertebrate remains showed breakage caused by the explosion, and fortunately all of these were labelled as such.

Very little damage was caused to the vertebrate remains following excavation. The large remains were washed carefully with a gentle detergent and water before conservation to remove adhering particles of clay and sand. Following this an attempt was made to remove some of the surface mineral crusts on display material, 20% of the cetiosaur remains showed fine preparation scrapes to remove these crusts, but these were easily distinguished from other scratches, such as trample marks and tooth marks, as expose a much lighter coloured interior than the earlier marks, are free of matrix and tend to be concentrated. Very little conservation work was needed on the majority of the cetiosaur material, and effort was made to re-join pieces of bone separated by large scale transverse fracturing and this was augmented by the liberal application of PVA or paraloid adhesive. Some of the more damaged bones were impregnated with a paraloid solution in acetone, these included waterlogged bones which had the appearence and pre-impregnation texture of 'soggy biscuit' (R. Vaughan, pers. comm., 1991). Although, this conservation work was essential to retain the bones, unfortunately it rendered the bones useless for taphonomic investigation, as they were heavily varnished with the paraloid and features such as abrasion and weathering could not be determined.

The microvertebrate remains to my knowledge were not affected by the explosion or excavation techniques employed on site. The static sieving techniques employed to extract the material from the enclosing sediment is also extremely gentle in comparison to more usual agitated sieving methods employed. The sediment and bone material encased within it, remain stationary in the tray, limiting damage to fragile specimens. However, some breakages were inferred from closely associated fragments of a few specimens found during the sorting of the residues (Metcalf & Walker, 1993). A (very) few specimens were broken (Fig. 10.19e) or lost during sorting and investigation of the assemblage. Specimens broken during extraction or sorting were re-assembled with paraloid in acetone adhesive.

Generally, the microvertebrate remains were not treated after extraction from the clays, although some were washed gently with soap and water. However, were limestone clasts was extremely abundant in the sieved residues, treatment with 10% acetic or formic acid was used to extract more vertebrate material. The acid was buffered with calcium orthophosphate or spent acid and the extracted remains observed for any damage by acid attack. A controlled experiment was also initiated to study the effects of preparation acid digestion upon fossilised bone and tooth material, in order to eliminate such alteration during taphonomic investigation. Initially 22 similar-sized crocodile teeth (0.5-1.0cm in length) from the Hornsleasow assemblage, in various

degrees of preservation, were placed in measured amounts of acetic acid (5ml) - the main acid used during the Bristol phase of microvertebrate extraction. Concentrations of acid used were :

1. 5% acetic acid - unspent and unbuffered.
2. 5% acetic acid with buffer-calcium orthophosphate ($Ca_3(PO_4)_6$).
3. 5% acetic acid - spent (used by Ward(1981) as a buffer).
4. 1/2 and 1/2 mixture of spent and unspent acetic acid (Ward 1981).
5. 5% nitric acid - control
6. Purite water - control

As acid attack may take place preferentially over various parts of the teeth, the specimens used were of two specific types, those in excellent condition without any imperfections and those of poor taphonomic condition, with imperfections such as cracked enamel, breakages and wear facets. The specimens were carefully cleaned (to remove dust and adhering clay/limestone particles) and washed with acetone, to remove any P.V.A. or paraloid used to repair or enhance specimens, and then finally cleaned in an ultrasonic bath for 5 minutes. Their dry weights and appearence was finally checked and the specimens placed within stoppered test tubes containing the solutions. Weight measurements and a microscopic study of the surface of each specimen was carried out at intervals of one hour, two hours, four hours, eight hours, twelve hours and twenty-four hours. After this the specimens were left in the solution for one week, being checked for modification or weight loss once every 48 hours. Any significant change in appearence or weight was noted and drawn.

A filter paper cradle was employed to hold the tooth within the acid, facilitating recovery of the tooth from the spent acid after each run. This ensured that handling of brittle demineralised teeth would be to a minimum, reducing the amount of finger grease to the surface (important before dry-weighing, and minimising the use of acetone cleaning fluids), and that broken pieces would remain upon the filter paper. Acetone cleaning was also kept to a minimum, as I noticed a white residue appearing upon the teeth after 2-4 hours digestion, this was suspected to be an acetone compound, calcium acetate, and in order to eliminate the risk of coating I reduced the washes to one per tooth, and increased the demineralised water rinses to three. Any mounting with blu-tak, or paraloid consolidant, needed a more careful cleaning procedure, which included manual brushing the surface with a wet or acetoned 0000 brush, before the washings. For visual mounting a sellotape screen overlying a blob of blu-tak gives an adequate and pliable base, and any residue is water soluble.

The results were extremely promising, of all the teeth (including those in a poor preservational state) only those within unspent acetic or nitric acid showed any sign of acid digestion, in the form of pitting of surfaces (Fig. 10.12f) and general flaking of the mineralised enamel. However, this was seen fairly quickly after initial immersion (2-4

hours), and hence, it is suggested that during any acid extraction, buffered solutions should always be used. At Gloucester Museum, formic acid has been employed for this procedure and is apparently much kinder to fossil vertebrate matter, although more difficult to handle (D.Dartnall and D. Ward, pers. comm., 1993).

The microvertebrate material curated at Gloucester Museum was stuck to card by a water-based paste. Although fine for the majority of specimens, the action of removing specimens from the cards for study, by soaking in water, caused fragile, porous or weathered material to loose flakes of surface or in very severe cases, fragment completely. Curation into tubes or loose in comparmented boxes is recommended for future studies.

10.8.2. Problems and taphonomic bias associated with the sampling methods

The original excavation in 1987-1988 set out to exploit as much of the clay lens for fossil remains as was possible to remove from the quarry. Every effort was made at that stage to correctly locate and describe the potentially fossiliferous sediment before removal to Gloucester City Museum for sieving. The careful labelling of each bag of clay included a site reference (on the gridded system) and clay context reference (Darlington, 1989). This enabled a detailed record of stratigraphic and spatial information for the recovered material during the following sieving projects. However, the problem with such a large scale excavation and sampling procedure is that it creates an incredible amount of duplicated data and information. This was inevitable when there was so many people working upon a single project, including excavators (archaeologists, museum staff and amateurs), technical and academic staff. In this project people tended to develop their own procedures and labelling scheme as they went along, hence creating a mountain of cross-referenced data to be studied. For instance, large bone or wood material was given an initial field label and description by the excavators, a PCF number for the conservation project, a temporary museum number (for the exhibition) and finally a permanent museum number. Another case of over-zealous and confusing labelling is the field context descriptions given by the archaeologists for the clays, another field description by R.F. Vaughan, and lastly HQ context numbers given during the sieving project to identify the location on the gridded system. Clearly, detailed labelling is extremely important in such a project, but people need to talk to one another and plan this out prior to the investigation.

The sieving procedure employed at Gloucester Museum involved pre-weighing the clay before bulk processing. The extracted remains were then sorted onto pre-printed curation cards, which meant that details upon the clay context sieved, its weight, the grid square sampled and treatments involved could be recorded on them. Ideally, details that are entered upon the cards should also include the card number, the technician's identity, site information, date, bag or batch number and any further

comments. These details and identification of the specimen were then catalogued on PC FILE database at Gloucester City Museum. All the remains from a bag of clay, regardless of taxonomic type were curated onto the same card, and this is to be recommended as an aid to subsequent taphonomic or palaeoecological investigations. Small problems have occurred in this technique when inexperienced samplers did not fully label the cards, and modern contamination (e.g. plant seeds and invertebrates) were sorted. As each pre-weighed bag of clay has a clay context and grid square number, the stratigraphic span and spatial derivation of the recovered specimens can be ascertained. Therefore even though the grid squares and the weight of clay sieved are arbitrary measures, taphonomic information concerning the degree of clustering in the assemblage, possible associated elements and the fossiliferous potential for each context can be ascertained. Obviously three dimensional spatial derivation (within paleosol horizon) will be lost during sieving, but this is not considered to be too much of a problem.

However, when the sieving project was transferred to Bristol University, this procedure was abandoned as unnecessarily complicated, as the onus was upon recovery of as much taxonomically viable vertebrate material as possible. It was thought that the assemblage might rival those recovered from over Bathonian sites (e.g. Kirtlington) in the amount and preservation of unusual or unique taxa. Alas, this was not to be the case, although the Hornsleasow assemblage contains Middle Jurassic mammals, amphibians and dinosaurs, they are by no means complete, perhaps too poorly preserved to be identified and so are extremely rare. For the 12 tonnes of sediment processed only a handful of these unusual taxa were identified (Chapter 9). Hence, Hornsleasow is taxonomically a relatively uninteresting site; a crocodile pond with an associated fauna of lepidotid fishes and pleurnosternid turtles. However, taphonomically the site had incredible potential.

Unfortunately the means of sorting at Bristol has decreased this potential. Although uninteresting specimens were not discarded, they were sorted into boxes containing many bags worth of one context of clay. Hence although the stratigraphic range was still upheld (all the material at Bristol came from two contexts, 1007 and 1017; both Grey Clay), the spatial information was lost. Bags of clay were not pre-weighed and therefore information regarding the fossiliferous potential and taxonomic sorting for these two contexts was also not investigated. The sorting meant that if one wanted to assess the relative abundance of a certain taxa or skeletal element one would need to study all the remains taken from one context, rather than a located and pre-weighed controlled sample. In the case of crocodile teeth or unidentified bone fragments this would have meant sorting through several thousand specimens - hence in this frame the method has cost more time than it saved and I have tended to rely upon

the Gloucester material, creating a bias for my taphonomic studies towards the Grey Clay assemblage (section 7.1).

Obviously when dealing with such a large load of fossiliferous sediment (20 tonnes of clay were removed in all), it is necessary to cut corners to avoid the research carrying on for decades. Also there must be a degree of sorting of the sample, otherwise the collection would become chaotic and unwieldy. However, it would be more manageable to carefully sample a controlled and pre-weighed measure of each context of clay for each or a small number of designated grid squares. After all, the excavators provided us with this information, therefore we might as well have put it to some use. This controlled sampling would in fact benefit not only taphonomic investigations, but the sieving project as a whole. Such a method could enable an investigator to have a prior knowledge about the batch of sediment to be processed - whether it contained rare specimens, potentially associated or even articulated material, and could save time in the long run. At Gloucester, the unsorted loads stuck to the cards occasionally produced such specialities, these were then removed from the cards (and a label attached to say so) and placed into numbered tubes for examination by experts. Obviously after the controlled sampling, one could continue the extraction and sorting process as was used at Bristol.

10.9. Macrovertebrate assemblage taphonomic history

The macrovertebrate material is largely the *in situ* scavenged and trampled remains of a single cetiosaur dinosaur. These remains underwent significant taphonomic modification before being buried (Fig. 10.20): the carcass was pulled apart by scavengers, trampled by other animals, weathered and abraded within the karstic hollow. After the stabilising effect of more paleosol deposition over the remains, the bones were mineralised and slightly crushed.

10.10. Microvertebrate assemblage taphonomic history

The taphonomic history of the microvertebrate assemblage is much more difficult to define. The remains probably accumulated in the hollow a combination of normal non-biological attrition (Maas, 1984) and as a result of coprocoenosis (Mellet, 1974). Studies of modern carnivore activity upon small prey (Dodson & Wexlar, 1979; Fisher, 1981; Korth, 1979; Mayhew, 1977; Mellet, 1974) have shown that different predators can produce quite separate scatological accumulations, and thus, may bias the sample according to dietary habits and/or digestion processes (Maas, 1984). For example, modern crocodilians almost completely digest the remains of their prey, leaving teeth as the only identifiable elements within their faecal pellets (Fisher, 1981), whereas some owls (Dodson & Wexlar, 1979; Korth, 1979) devour their prey whole, and many elements (teeth, mandibles, limbs) can be recovered. Korth (1979) concluded that if a

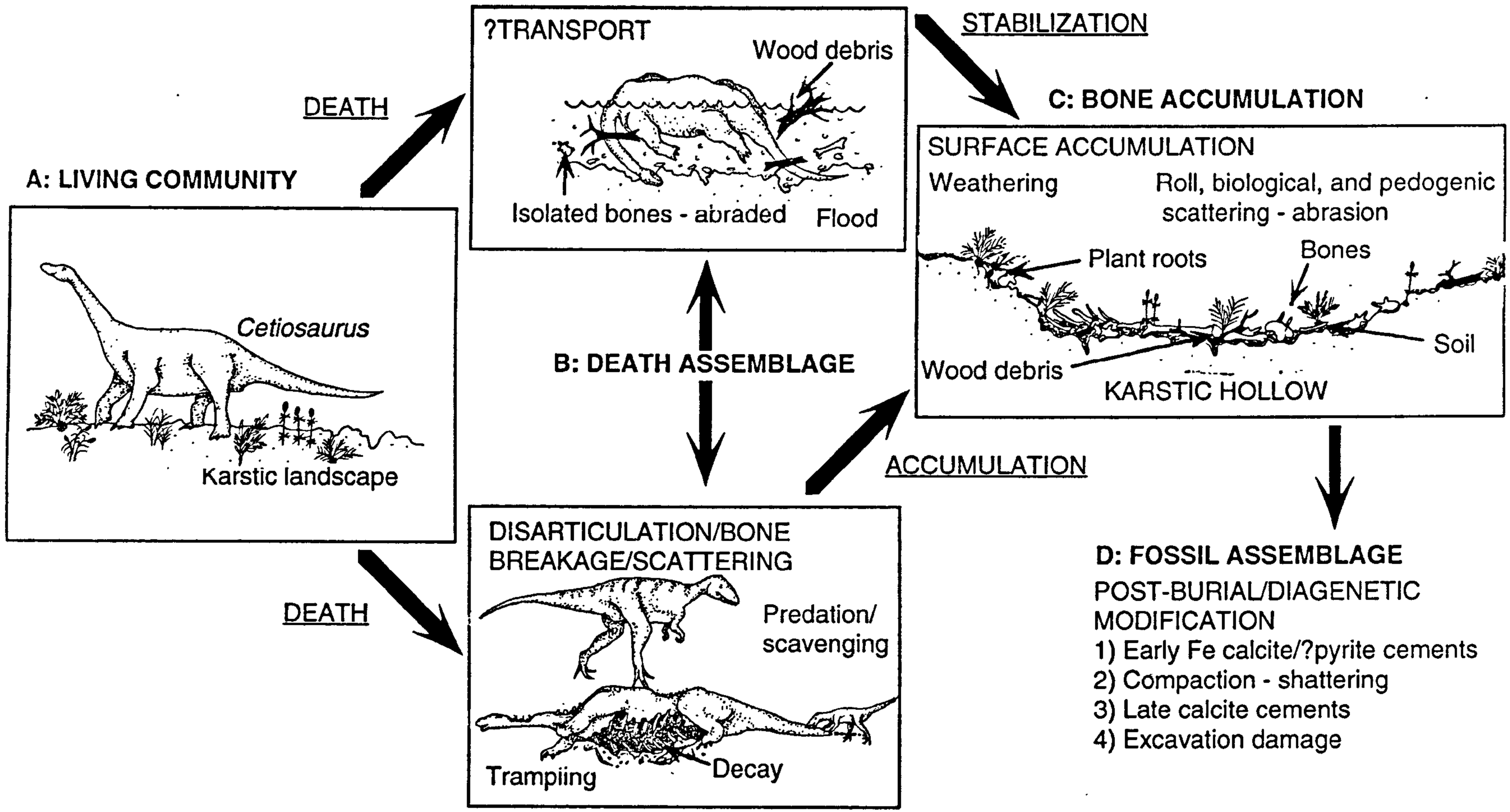


Figure 10.20. Proposed taphonomic history for the *Cetiosaurus* skeleton recovered from Hornsleasow Quarry

microvertebrate accumulation had been caused by coprocoenosis then there should be a fairly high proportion of elements represented for each animal identified in the concentration and this is certainly true for some of the Hornsleasow indigenous fauna, such as fishes and crocodiles. Korth (1979) also stated that in mammalian and raptorial bird scatological assemblages, the majority of limb-bones and skulls are unbroken and common and that the remains of smaller animals are less likely to be fragmented, whilst those of larger animals may show tooth marks. Evidence for this is rather scant in the Hornsleasow assemblage, but this might be due to later taphonomic modifications. Clearly, the microvertebrate assemblage at Hornsleasow contains the remains of animals which have accumulated from coprocoenosis.

As a watering hole with an associated dense vegetation, the Hornsleasow pond would make an attractive location for herbivores and predators alike. Modern watering holes contain dense accumulations of bones: the remains of multiple individuals which may have died due to thirst, starvation and may attract predators (Haynes, 1988, 1991) and are said to serve as "magnets drawing in nutritionally stressed individuals or potential prey" (Lyman, 1994, p.193). Therefore, remains represent indigenous aquatic animals and their prey. Attritional bone assemblages would be expected in such a situation and periods of low sedimentation (e.g. Grey Clay times) could result in high bone concentration on and in the soils there.

Some exotic specimens (e.g. large dinosaur teeth) were no doubt transported to the site during depositional flooding episodes. As a topographic low, the pond would be likely to receive and protect organic remains and localised transport, sorting and hydraulic concentration of bones in this environment might occur from sporadic flooding (cf. Behrensmeyer, 1988). Microvertebrate remains would accumulate as alternating periods of passive accumulation by attritional means, and relatively rapid sedimentary deposition from flooding that could bury bone rich horizons (Behrensmeyer, 1988). Trampling by animals within this environment would contribute to disarticulation and the breakage of the bones, but would also enhance their burial in soft substrate of the clay soils.

10.11. Palaeoecological factors of the Hornsleasow vertebrate assemblage

The palaeoecology of the Hornsleasow assemblage is briefly outlined below. The assemblage is considered as a whole, but with emphasis on the sieved fractions (i.e. microvertebrate component) rather than the single cetiosaur skeleton.

10.11.1. Time averaging

Fossil assemblages that have formed by multiple episodes of attrition are termed time averaged (Behrensmeyer, 1982). Clearly, the Hornsleasow vertebrate assemblage derived from the paleosol units is temporally as well as ecologically mixed (i.e. a

'mixed' assemblage *sensu* Kidwell & Bosence, 1991). In this sense then the vertebrate accumulation is similar to the invertebrate assemblage and much of the arguments for time averaging of that component (section 8.2.4) is applicable and shall not be repeated here. In conclusion, the vertebrate assemblage within each paleosol unit has been time-averaged over a period of between 10^2-10^5 years.

10.11.2. General palaeoecology

Maas (1984) has speculated upon the use of microvertebrates in palaeoecological reconstructions, and concludes that pre-burial forces are most effective at destroying palaeoecological information. She suggests that although this initial loss of information will bias the relative abundances of individuals within a palaeocommunity, it will not affect actual diversity or 'species richness' (Maas, 1984, p.123). As predators often show a distinct preference for certain prey types, any accumulation which has had scatological input from several different species of carnivore should show a relatively good approximation to a living faunal community (Korth, 1979; Worthy & Holdaway, 1994). Therefore excepting the component derived from non-biological attrition, a probable palaeocommunity for the Hornsleasow assemblage can be estimated.

The predominance of non-marine aquatic forms is striking in the assemblage and this reflects the overall palaeoenvironmental reconstruction of the clays as a pond environment. Obviously the fish, turtle and crocodilian components were the most important in terms of abundance and were almost certainly indigenous to the pool. Of the fish *Polyacrodus* is an euryhaline taxa (Cappetta, 1987), with freshwater examples from Purbeck Beds of Dorset (Upper Jurassic; Woodward, 1916) and the Wealden deposits of southeast England (Lower Cretaceous; Patterson, 1966) and *Lepidotes* material has also been recovered from these and other non-marine horizons. Similar pleurnosternid turtle and goniopholid crocodile remains have been recovered from the Kirtlington Quarry microvertebrate accumulation, as has amphibian material. The impersistent *fimbriata-waltoni* clay from which the vertebrates are derived at Kirtlington is considered to represent deposition into a brackish or fresh-water lagoon (Evans & Milner, 1994). The rarer non-aquatic elements in the assemblage represent the floodplain community, but little can be said about their palaeocommunity structure.

Chapter Eleven Conclusions and Implications

11.1. General conclusions: answering the specific research aims

The rationale behind the research written about in this thesis was to provide a thorough description of a newly excavated palaeontological site. The original intention of the excavators and museum staff was to recover as much material and information from the site in order that such an investigation could be made. When the project transferred to Bristol under the direction of Dr. Benton, it was intended that I should carry out a palaeontological study of the microvertebrate remains recovered by the sieving. The thinking behind this lay in the wake of the set of palaeontological investigations into the Upper Bathonian 'Kirtlington Mammal Bed' microvertebrate horizon from Kirtlington Cement Works, Oxfordshire by a team from University College, London (UCL) in the 1970-80's. The publications which arose from the UCL research concerned the reconstruction of specific components of the Kirtlington fauna, rather than a general overview of the palaeoenvironment and ancient ecology of the assemblage as a whole. These important papers served to enlighten the palaeontological world upon a little known period in the history of terrestrial vertebrate evolution during the Middle Jurassic. They included a review of the important mammal components (Kermack, 1988; Kermack *et al.*, 1987), identifications of the earliest modern frog (Evans *et al.*, 1990), the oldest known salamanders (Evans *et al.*, 1988), the first Jurassic choristodere (Evans, 1989, 1990), and the first true lizards (Evans, 1991). The site has also yielded one of the last occurrences of the tritylodont mammal-like reptiles (Freeman, 1976, 1979). The painstaking palaeobiological reconstruction work by the UCL team continues to this day, with the result that they are shedding light on the composition of the Bathonian palaeocommunity structure.

It was hoped that the Hornsleasow fauna could also provide new and unique vertebrate material, and that the boundaries of the evolutionary history for these groups of vertebrates could be pushed back even further in time: the Hornsleasow assemblage being some six ammonite zones older than the Kirtlington site. However, although initial progress seemed to indicate that this might be the case, with the similar amphibian, reptile and mammalian taxa being recovered from the sieving, after further research I realised that I could not reproduce the same calibre of palaeobiological work as that initiated by the UCL team. The reasons for this discrepancy are two-fold, firstly, as a geologist I have approached the task through a completely different viewpoint - at UCL, all the sediment was discarded after washing for vertebrate remains (A. Milner, pers.

comm.) - a fact which to my geologist ears was horrifying, and any material considered to be 'rubbish' was similarly ignored in their studies, so that the true taphonomic environment of the site could not be made. Secondly, and more importantly, the material coming out of the Hornsleasow project was much less well-preserved. Much of our material as documented in Chapter 10, was incredibly broken and undiagnostic, and hence, any taxonomic investigation was hampered from the start. What good material was found tended to either confirm Kirtlington taxa in the assemblage (e.g. *Cteniogenys oxoniensis*, 'Kirtlington Salamander A', etc.: Chapter 9), rather than be totally unique to the site. No new taxa were described from the Hornsleasow project over its eight year research initiative.

However, as the preceding 400 pages have shown, there was much that could be gleaned from the project and in doing so, I hoped to carve out a new niche for myself in the realms of Bathonian palaeontology. The three specific aims of my research were outlined in Chapter 1, they are:

1) To combine depositional environmental studies with preservational factors for coastal-plain assemblages;
2) To review the state of vertebrate taphonomy for non-mammalian microvertebrate assemblages;
3) To review the palaeoecology of the British Middle Jurassic vertebrate palaeocommunities.

In the following sections I shall attempt to defend these aims with some very generalistic concluding points that can be made from my thesis research. In doing so I shall be drawing upon research which I undertook during my postgraduate studies on other Bathonian sites, but which was not referred to in the thesis (through considerations of space and time), in order to illustrate how Hornsleasow compares and whether such a study could be undertaken on other horizons. I hope to expand on my research in order to speculate on the broader implications of the research for future palaeontological investigations.

11.2. Specific Research Aim 1: Palaeoenvironmental conclusions

To summarise the general conclusions made at the end of Part B (Chapters 3-6), the Hornsleasow microvertebrate-bearing horizon represented deposition of clay grade sediment into a free-draining hollow, which had developed during coastal karstification of a marine limestone in a period of emergence in the earliest Bathonian (*zigzag* Zone: Chapters 3-4). The clay infill was probably introduced during a flooding event, and the source seems to be a mixed terrigenous clay and weathered ash deposit (Chapters 5-6). Following blockage of the drainage system at its base, the hollow became a coastal-marsh pond and supported a diverse freshwater assemblage of plant and animal life (Part C). Later, the pond became silted up, and development of a fairly well-drained paleosol

was initiated (Chapter 5). This in turn was followed by marine transgression, and the clay infill was covered by a beach-like deposit of carbonate sands. After which a return to marine limestone deposition followed (Chapter 5), marking the end of this brief terrestrial interlude in Chipping Norton Limestone Formation sedimentation.

The implications of such a deposit being found within the Bathonian succession in the English Midlands are discussed below, and speculations as to the wider importance of the palaeoenvironmental conclusions reached by this thesis are discussed.

11.2.1 Wider implications of the palaeoenvironment of the English Midlands in the Bathonian

The justification for the detailed nature of the palaeoenvironmental work which I carried out during my research is that Hornsleasow is essentially a fully terrestrial site in the earliest Bathonian, and at a palaeogeographic position which was some 60km from the nearest unequivocal palaeoshoreline (the London-Brabant massif to the east, although the conjectural Pennine landmass may have been somewhat nearer). At the beginning of the excavation and research, several scientists argued that the terrestrial clays and their enclosed fauna had been transported into a marine carbonate shelf environment during a storm event, and dumped *en masse*. Therefore any conclusions made from the work would only be of a general interest in the context of the ancient environments of the Cotswolds-Weald shelf as a whole. Henceforth, I set out to prove that the horizon was not only terrestrial in origin, but also *in situ*.

The proof that this is the case has several broader implications for the palaeoenvironmental reconstructions of the English Midlands in the Early Bathonian, suggesting that sea-level had fallen enough for part of the shelf to be exposed. The evidence ties in with the knowledge that during much of the Lower Jurassic the Cotswolds region was subjected to uplift and subsidence in relation to a series of north-south trending horst and graben structures of Late Triassic initiation which underlie the area (e.g. Chidlaw, 1987). Their tectonic control upon sedimentation patterns in the English Midlands may have continued into the early part of the Middle Jurassic, and that the Hornsleasow paleosol lies upon the flank of one of the major horst structures (Moreton-in-Marsh 'swell') suggests that uplift may have also occurred at this time.

The evidence for volcaniclastic material in the fossil soils (Chapter 6) and elsewhere in Britain at the same time (specifically the Fullers Earth bentonites of south-west Britain and the volcanic doming in the Forties-Piper region of the North Sea) indicates that major tectonic events were happening in the British section of the North Atlantic rift system. The tectonism seems to have begun in earlier Mid Jurassic times (Aalenian-Bajocian) with regional upwarping of the crust in the Bristol Channel region, around the edges of the London-Brabant massif, the Cleveland Basin and the Hebrides area. It is most likely that this uplift accounts for the extremely shallow nature of the

marine sedimentation in the English Midlands at the time, leading to the development of the Bahaman-type carbonate platform regime (Great Oolite Group) which would have been prone to regional regression and subaerial exposure. The presence of other freshwater deposits in the Great Oolite Group of the Cotswolds, such as the lagoonal sediments of the Sharps Hill Formation, a coastal marsh and probable karst in the Hampen Marly Formation, and freshwater lagoonal deposits in the Forest Marble Formation, indicate that this was the case throughout the Bathonian. More research of the type outlined in this thesis on these horizons and a search throughout the succession for more of the same, could throw light upon the structural controls and sedimentation patterns in the British Bathonian. Such a field study might also produce many more terrestrial vertebrate-bearing units.

The elucidation of the palaeoenvironment for the Hornsleasow paleosol horizon also provided palaeoclimatic information for the British Bathonian succession. It suggested a hot and fairly humid climate which was prone to periodic flooding and storm events. This information fits in well with the suggestion that at this time, the British Isles lay within an area subjected to seasonal monsoonal weather, although during the rest of the year the climate would have been hot and reasonably arid. Evidence for this palaeoclimate pattern in northern Europe is fairly scant, because of the predominance of marine rocks, although the commonly held belief that the Tethyan and Boreal Realms may have responded to a palaeotemperature gradient in the seawater has led to some climatic information to be gathered. However, similar faunal provincialism is not seen in the Bathonian terrestrial faunas between the southern British region and the localities in the Hebrides, as all components of the Great Oolite Group assemblages have been recorded recently from the Kilmaluag Formation of the Great Estuarine Group on Skye (Evans and Waldman, 1994). No attempt has been made in any of the other Bathonian terrestrial-bearing localities to look for palaeoclimatic information, and it is my aim to carry out more work on this subject in the future. I hope to test the information gleaned from traditional palaeontological and sedimentary means against oxygen isotopic methods of palaeotemperature determination on fossilised bone and carbonates in these deposits. If this proves fruitful, the method could prove useful in determining palaeotemperature measurements for fossils within the epicontinental marine rocks of northern Europe, in order that a truer palaeoclimatic model can be made for the whole region.

11.2.2. What can be gained from undertaking a similar palaeoenvironmental analysis

As well as providing the useful information on the nature of the sedimentary sequence in that area, such a palaeoenvironmental study can help immensely in the discussion of the fauna contained therein. Indeed, it is my hope that detailing the prevailing conditions in which so much vertebrate and non-vertebrate fossil material was recovered, could prove

useful for ascertaining potential non-marine fossiliferous horizons. Paleosols are not usually thought of as exceptionally rich in fossil material compared with other non-marine settings (e.g. river sediments), and their recovered assemblages are not especially useful for taphonomic or taxonomic investigations. However as this study shows, in the absence of other terrestrial deposits, they are worth sampling. Of course the Hornsleasow paleosols developed at first within a small pond in a coastal marsh setting (a much more attractive environment for animal life), and the taphonomic conditions seem to have been unusual for so much material to be preserved. However, similar sieving explorations into what appear to be blank, structureless horizons, or as Channon (1950, p. 249) described the first Hornsleasow clay lens "barren of all life", would probably reveal more paleosol biotas.

11.2.3. Future palaeoenvironmental research in the Mid Jurassic

In general, I believe that this type of careful palaeoenvironmental research could be carried out to some degree on most of the previously described vertebrate-bearing sites in the British Bathonian sedimentary sequence (section 11.4) and any new sites. The general procedures which should be adhered to in carrying out such work are laid out in the thesis and can be summarised as follows:

(1) Gather as much information about the fossiliferous horizon in the field. This is common sense, as this will influence all the future research, and once a site has been worked out there is no real way of reconstructing the environment without detailed field notes and drawings. It is important at that time to have everything which is removed from the site carefully labelled, and photographed or drawn, so that its position on site can be determined. Samples of sediment should be taken as well, and a clear log of the section made with their location indicated.

(2) Once all the material is back at the laboratory, the researcher should take a step back from the information gathered in the field and decide upon a clear set of procedures which will yield the best overall palaeoenvironmental results. This generalistic approach should cut out the time and sample wastage which might occur if the researcher were to plunge into research upon specific parts of the project.

(3) The best method of gleaning information from a site is to take a 'step-by-step' approach to the research. By answering each question in turn, this technique will provide the clearest interpretation, and in doing so, should lead the researcher on into the next field. I applied this method, to some degree of success to both the determination of the environment of formation of the paleosol and to the taphonomy of the fossil assemblage.

Of course this procedure is common sense, and could be applied to other sedimentary regimes as well. No such detailed investigation was carried out on any of the other vertebrate-bearing sites in the Cotswolds Bathonian and thus, only speculations

as to the environmental preferences and ecology of the animals which have been described from them can be inferred. Most of these accounts cannot even say with confidence whether the material have been transported to its present resting place. However these factors could be re-interpreted for the previously described sites and there faunal assemblages (sections 11.4-11.8).

11.3. Palaeoecology and taphonomy of the Hornsleasow biota - general conclusions

The Hornsleasow biota can be split into three mutually exclusive groups, based upon their palaeoecological and taphonomic characteristics, and these can be summarised as follows:

(1) *Indigenous biota.* This group consists of the plants and animals which lived in and around the pond environment, and includes a low diversity and low abundance epifaunal invertebrate community of freshwater gastropods, and a vertebrate fauna of aquatic or semi-aquatic forms, such as the fish, amphibians, crocodiles, turtles and champsosaurs. The indigenous microvertebrate assemblage is somewhat complicated by the addition of terrestrial forms (from predation and coprocoenosis around the watering hole, tooth loss and attrition by normal depositional processes) which were contemporary with those living in the pool. The plant biota comprises the rootlets of lycopods and ferns. The biota supports the sedimentary evidence that the clay paleosols represented sedimentation in a low-nutrient freshwater pond environment with brackish influences, and which eventually silted up.

The preservation of these forms indicates that the pond had an alkaline environment, with basal anoxia in the lower waters and the underlying Grey Clay soil. Corrosion damage to shells and vertebrate fossils indicates that in places the porewaters were acidic.

(2) *Contemporary derived biota.* As mentioned above the biotic assemblage is ecologically mixed, although not time-averaged. Storm washover would have introduced marine invertebrates from marginal settings, whilst flooding on the surrounding karstic plain would have brought into the hollow, skeletal remains picked up from surface deposits, as well as the remains of forms killed by the actual floodwaters. For example, such an exotic element is the cetiosaur dinosaur skeleton which underwent disarticulation and some disassociation before being swept into the karstic hollow. Other such exotic forms include wind-borne spores and pollen, and waterborne plant fragments, including charcoal.

The contemporary derived material was as equally important as the indigenous biota, as it provided information on the floodplain as a whole.

(3) *Time-averaged material.* This group consists of exotic skeletal fragments derived from the karstic weathering of the Chipping Norton Formation limestones. Their presence can add nothing to the ecology of the pond/floodplain, but did provide an

insight into palaeoecology of limestones, and much information upon subaerial weathering of bioclastic clasts.

11.3.1. Implications for similar studies

A similar step-wise approach was undertaken for the research into the palaeontology of the Hornsleasow biotic assemblage, as for the geology. The initial description of the components of the accumulation formed the bulk of Chapters 8-9, and then proceeding with the interpretation of their associations in Chapter 10. This was important as Hornsleasow represents a unique snapshot of life in Bathonian times, and the documentation of the fauna is an important result by itself. The accumulation had also been sorted by ecological, depositional and post-depositional forces, hence the study provided an insight into the nature of this sorting in providing a taphonomic history of the deposit and a probable palaeoecological model for the palaeocommunities represented (Chapter 10). The study has showed just how much palaeoecological information can be determined from a single biotic assemblage, and the need for integration of the palaeontology of non-vertebrate and vertebrate fossil material with sedimentary evidence.

The determination of a specific taphonomic history for the whole assemblage was facilitated by recognising the specific components of the accumulation, i.e. in terms of taxonomic groupings (vertebrates, invertebrates and plant fossils), morphological characteristics (e.g. variations in response to preservation for different size of vertebrate material), preservational features (e.g. mineralisation) and palaeoecological considerations (e.g. derived, time averaged or indigenous fossils). Once a taphonomic pathway was determined for these specific groups, a true picture of the palaeoecology of the site was resolved - a pond assemblage with indigenous biota affected by depositional and post-depositional taphonomic overprints. From this study I was able to confidently state which elements were *in situ* and which were not, in order to produce a palaeocommunity structure for a Middle Jurassic terrestrial accumulation.

The mixed nature of the assemblage was also crucial as one could also gain an insight into more than the palaeoecology of the pond environment, but of the whole surrounding floodplain. For example, stegosaur dinosaurs clearly did not inhabit small karstic ponds, but the relative abundance of their teeth in the assemblage compared to other large herbivorous dinosaurs (e.g. *Cetiosaurus*) suggests that they were a common component of the Cotswolds coastal plain environment. The windborne palynofloral debris also showed that the coastal plain was vegetated in a lush mixed fern and lycopodaceous flora, with drier areas covered in conifers, cycads and ginkgoes. Whilst the presence of charcoal in the soils suggested that this area was prone to wild fire ignition.

	Hornsleasow	Kirtlington Mammal Bed	Swyre/Watton Cliff	Eyford
Chondrichthyes				
marine	-	+	+++	++
freshwater	+	+	-	-
Osteichthyes				
marine	-	+	++	++
freshwater	++	++	((+))	-
Amphibia				
freshwater	((+))	++	(+)	-
Reptilia				
aquatic (marine)	-	((+))	-	++
aquatic (freshwater)	+++	+++	+	-
airborne	++	++	+	+
terrestrial	++	++	+	(+)
Mammalia				
terrestrial	(+)	+	(+)	-
Invertebrates				
marine	+-	(+)	+++	++
freshwater	+	+	-	-
Plants				
in situ	+	-	-	-
derived	++	++	((+))	+-
Sedimentary environment	Pond/Marsh (freshwater)	Lagoon (freshwater)	Subtidal shell bank/channel	Subtidal storm beds/channels
Taphonomy				
weathering	+-	+-	-	-
abrasion	+-	(+)	++	++
fragmentation	+++	(+)	+	+
digestion/TM/GM	++	(+)	-	-
corrosion	(+)	-	+	+ (bio.)

Table 11.1. A comparison in the composition of the biotas, palaeoecologies, sedimentary and taphonomic environments of Hornsleasow, Kirtlington, Watton Cliff/Swyre and Eyford microvertebrate-bearing facies.

Relative abundances: +++ = very common; ++ = common; + = present; (+) = uncommon; ((+)) = rare; - = very rare or absent; +- = highly variable.

TM = trample marks; GM = gnaw or bite marks; bio. = biocrosion

11.4. Specific Research Aim 2: Comparisons with other Bathonian microvertebrate sites

A similar study, including integrated sedimentary and palaeontological investigation could elucidate for some of the previously described terrestrial fossil-bearing Bathonian sites whether their faunas represent actual palaeocommunities or mixed allochthonous assemblages. Over forty vertebrate-bearing sites in the British Bathonian succession have previously been found, although only a few of these have been studied in detail. In order to provide a neat comparison of the taphonomic conclusions reached for Hornsleasow, I have studied three other terrestrial-vertebrate producing sites from the Great Oolite Group of southern and central England. This has taken the approach of a literature search (references indicated), limited fieldwork (except in the third case: section 11.7), and a taphonomic investigation on what fossil material (vertebrate and invertebrate) I could study. Table 11.1. illustrates the basic differences in terms of (known) sedimentary environment, biotic composition and taphonomy. The accounts will also highlight the many problems involved in working on non-mammalian microvertebrate assemblages (research aim 3).

11.5. ?Freshwater lagoonal palaeoenvironment: Kirtlington Old Cement Works Quarry

The old cement works quarry at Kirtlington, Oxfordshire (SP 494199) exposes a section of Mid-Late Bathonian age through the White Limestone Formation, Forest Marble Formation and Lower Cornbrash. Fossil fishes, amphibians, reptiles, and mammals have been collected in recent years from the *fimbriatus-waltoni* Beds in the White Limestone Formation and from the Kirtlington Mammal Bed, a microvertebrate locality near the base of the Forest Marble Formation. The vertebrate faunas have been described by numerous authors including the original collector E. Freeman (1976, 1979) and the UCL team: Evans (1989, 1990, 1991, 1992), Evans *et al.* (1988, 1990), Evans and Milner (1991, 1994), Kermack (1988) and Kermack *et al.* (1987). Kirtlington has been designated an SSSI for its vertebrate assemblages, and is recorded in the Geological Conservation Reviews for fishes and amphibians (Dineley and Metcalf, in prep.) and reptiles (Benton and Spencer, 1995). The following account draws upon their work and a separate taphonomic study which I made upon material in the UCL collections in 1993.

11.5.1. Geological setting

The succession in the quarry has been described by Odling (1913, pp. 493, 494), Arkell (1931, pp. 570-2), Douglas and Arkell (1932, pp. 123, 124), and Richardson (1946, pp. 69-71, 78-79). Additional information has been provided by McKerrow *et al.* (1969) and Freeman (1979). The following composite section (from Benton and Spencer, 1995) is based on these authors, and Richardson *et al.* (1946), in particular,

with additions from Palmer (1973, 1979) and Torrens *in* Cope *et al.* (1980b, p. 36) (numbering of individual beds is from the top down):

Lower Cornbrash	2.97m
Forest Marble Formation	
1. Clay, grey and buff, with some thin, irregular hard bands	1.53m
2. Clay, dark grey (= beds 3w-z of Freeman 1979)	0.69m
3. Limestone, yellowish, flaggy, locally marly and 'shaley', oolitic, with occasional inclusions of white lithographic limestone; ripple marks, rain pits (? =bed 3v of McKerrow *et al.* 1969; Freeman 1979)	0.61 - 0.92m
(White Limestone Formation)	
4. Clay, grey-blue, with three pale mudstone layers, one at the bottom (= beds 3p-u of McKerrow *et al.* 1969; Freeman 1979; = Unfossiliferous Cream Cheese Bed of Odling 1913 and Arkell 1931). The basal unconsolidated 0.04-0.25m brown marl unit (Bed 3p) is the Kirtlington Mammal Bed of Freeman (1979)	2.0m
(White Limestone Formation)	
5. Coral - *Epithyris* Limestone (Upper *Epithyris* Bed or 'Fossiliferous Cream Cheese Bed' of Odling 1913 and Arkell 1931; ? Beds 3n-o of McKerrow 1969). Limestone; at northern end an extremely hard white blue-hearted lithographic rock. Passes locally into unfossiliferous oolite	1.22 - 2.21m
6. *fimbriatus-waltoni* Beds (= Bed 10 of Arkell 1931; Beds 3k, l of McKerrow *et al.* 1969). Clay, grey-green to greenish black, with some white pellets at top; bed largely made up of bivalves; when bed 7 is absent, there is a lignite at the base	1.07m
7. Oyster - *Epithyris* Marl (= Bed 9; Middle *Epithyris* Bed of Arkell 1931; Bed 3k of McKerrow *et al.* 1969). Marl, brown. Locally, a thin layer of corals occurs below	0 - 0.75m
8. Limestone, hard, blue-hearted (? = Beds 3i, j of McKerrow *et al.* 1969)	0.92m
9. Marl (? =Bed 3h of McKerrow *et al.* 1969)	0.23m
10. Limestone, similar to 8 (? =Bed 3g of McKerrow *et al.* 1969)	0.84 - 0.92m
11. *Epithyris* Limestone (=Lower *Epithyris* Bed of Arkell 1931; ? = Bed 3a-f, Bed 1e of McKerrow *et al.* 1969). Limestones, white, at west end of pit a mass of the brachiopod *Epithyris*. Thins out eastwards and replaced from beneath by lenticular limestones	2.44m

12. *Aphanoptyxis ardleyensis* Bed. Limestones, well bedded	0.46 - 0.61m
13. *Nerinea eudesii* Beds. Limestones in three courses	1.68m

Certain units, such as the *Epithyris* Limestone (Bed 11; Bed 1e of McKerrow *et al.* 1969), are laterally impersistent. There are considerable problems with correlating the lithostratigraphy of the units in this quarry with those elsewhere in the northern Cotswolds, and these particularly concern the placing of the boundary between the White Limestone and the Forest Marble. McKerrow *et al.* (1969) attempted a definition based largely on the occurrence of oysters, and took the basal bed of the Forest Marble to be at the base of the Oyster-*Epithyris* Marl (bed 7). Palmer (1973, p. 61) points out that at Kirtlington the Coral-*Epithyris* Limestone (bed 5) shows a typical White Limestone fauna and lithology, and he proposed that the Forest Marble/White Limestone boundary should be moved to between beds 4 and 5. This view was also expressed by Barker (1976) on the basis of a study of the gastropods. Palmer (1979) further argued this point and divided the White Limestone Formation into three members, of which the Ardley Member (beds 8-13) and the Bladon Member (beds 5-7) are seen at Kirtlington. Palmer (1979, p. 208, fig. 5) made it clear that his Bladon Member is intended to include both the *fimbriatus-waltoni* and Upper *Epithyris* Beds of the Cherwell valley which rest on the *A. bladonensis* Bed. In general, Torrens *in* Cope (1980b, p. 36) recommended that the base of the Forest Marble be taken as "the base of the clay overlying the Coral-*Epithyris* bed, or of the bed above at Kirtlington" (i.e. the base of bed 3 or 4).

In Bathonian times Kirtlington lay on or near the south-west shore of a small island barrier some 30km from the coast of the Anglo-Belgian landmass at a subtropical latitude of about 30°N (Palmer 1979), in the area known as the "Oxfordshire Shallows" (Chapter 6). The presence of abundant lignite, charophytes, and freshwater ostracods and gastropods in the marly sediments of the succession suggests a coastal environment, which had low relief, with creeks, lagoons, and freshwater lakes, rather like the Florida Everglades (Palmer 1979) and similar to that developed in early Bathonian times around Hornsleasow.

Vertebrates occur in two Impersistent horizons: the *fimbriatus-waltoni* Beds (base of the Bladon Member, Palmer 1979), and the Kirtlington Mammal Bed (Benton & Spencer, 1995). The composition of the Mammal Bed assemblage is shown in Table 11.1. The faunas of the two vertebrate-bearing beds at Kirtlington are rather different, which probably relates to preservational and environmental conditions, as well as sampling methods, rather than to the very slight age difference. They will be discussed separately.

11.5.2. The *fimbriatus-waltoni* Beds

These beds are made up of greenish-grey or black clay, which is often lignitic toward the base. Phillips (1871) recorded fish remains in association with the large bones of the sauropod dinosaur *Cetiosaurus oxoniensis*, within the lignite resting upon the eroded surface of the underlying limestone. Since then other authors have recorded large reptilian remains from the same beds including bones of the dinosaur *Megalosaurus* and teleosaur crocodiles (Richardson *et al.*, 1946; Arkell, 1931, Benton & Spencer, 1995).

The biostratigraphy of the Bathonian at Kirtlington is difficult since no ammonites have been found locally, and very few elsewhere in comparable rocks (Torrens 1969a, *in* Cope 1980b). However, the vertebrate-bearing *fimbriatus-waltoni* Beds (base of the Bladon Member) are dated as upper *hodsoni* Zone (basal Late Bathonian) (Torrens 1980b, p.36). Although, the occurrence of the ostracod *Glyptocythere penni* in the *fimbriatus-waltoni* Beds led Bate (1978) to suggest that this unit belongs to the *discus* Zone.

Palaeoenvironmental interpretations have been made on the basis of the sedimentology of the *fimbriatus-waltoni* Beds. McKerrow *et al.* (1969, pp. 61-64, 80) noted the abundance of lignite and occasional caliche-like nodules which they interpreted as indicating shallow water with occasional subaerial exposure. Although Palmer (1979, p. 210) regarded the calcareous clasts as pebbles formed by erosion of an incompletely cemented limestone bed. Palmer (1979, pp. 210-1) noted the complex channelled interdigitations of this unit at Shipton-on-Cherwell, Oxfordshire (SP 4717), and suggested that deposition of some of the clays was local and catastrophic, and that the nodules were derived from elsewhere. There is a non-sequence at the top of the *fimbriatus-waltoni* Beds, and localised emergence at this level is probable. Palmer (1979) proposed a quiet-water lagoonal environment subject to periodic current activity and influx of new sediment, perhaps during storms.

I have not made a study of any of the fossil remains recovered from the *fimbriatus-waltoni* Beds, although some palaeoecological and preservational information can be gleaned from the literature.

The only early reference to the vertebrate fauna of Kirtlington is a brief mention of "teeth of *Hybodus*, *Pycnodus* and *Strophodus*, and scales of *Lepidotus*" (Phillips, 1871, p.244) in a list of fossils of the Great Oolite in Oxfordshire. There is no clear reference as to whether the remains occur within *fimbriatus-waltoni* Beds or another horizon, but Phillips also lists reptile finds, which are considered to have come from this unit (Benton & Spencer, 1995). The fish fauna represents a marine assemblage, although Phillips does not figure any of the specimens, it is likely that the *Asteracanthus [Strophodus]* and *Hybodus* teeth are similar to species within other Great Oolite sections, such as those of the Stonesfield Slates. *Hybodus* is a ubiquitous element in the

classic Great Oolite sequences and the presence of this genus in the *fimbriatus-waltoni* Beds is not remarkable. The reference to both pycnodont teeth and semionotid scales (Phillips, 1871) is also consistent with a shallow marine fauna, similar in composition to other Great Oolite fish-bearing localities.

The tetrapod fauna is dominated by the long-snouted piscivorous crocodilians, *Steneosaurus* and *Teleosaurus* represented by their teeth, vertebrae and jaws. These were functionally adapted for an aquatic lifestyle, and probably inhabited a marginal marine environment. Their remains are relatively common in the Bathonian marine limestones of England and France. Plesiosaur remains have also been identified from the *fimbriatus-waltoni* Beds and would suggest marine conditions, but their rarity here may connect with a predominantly lagoonal/coastal marine situation.

Unusually, many skeletal remains (mainly vertebrae, limb elements and skull bones) of the terrestrial sauropod dinosaurs, *Cetiosaurus* and *Bothriospondylus*, are found in the *fimbriatus-waltoni* Beds at Kirtlington (Benton & Spencer, 1995). Another terrestrial element is a tooth of the meat-eating dinosaur, *Megalosaurus*. Their inclusion in the assemblage suggests a close proximity to land.

11.5.3. The Kirtlington Mammal Bed.

This horizon (bed 3p of McKerrow *et al.* 1969) was an impersistent lens, 21.5m long and 0.04-0.25m thick in the north-eastern corner of the quarry (Freeman 1979, p. 136). Sampled for vertebrate remains in the late 1970's and 1980's, it has almost certainly been worked out on the existing exposure, meaning that further recovery would only be possible by extending the present quarry face. The Kirtlington Mammal Bed falls within the *aspidoides* or *discus* Zone (Freeman 1979, p.136). The vertebrate fauna within this unit occurs as dissociated remains of a variety of bony fishes, sharks, amphibians, reptiles and mammals (Evans and Milner, 1991, 1994) (Table 11.1). Associated fossils include microscopic freshwater charophytes, indeterminate plant fragments, shell fragments and ostracods.

The contacts of the bed with the Coral-*Epithyris* Limestone below (bed 3o of McKerrow *et al.* 1969) and another limestone above (bed 3q) are extremely sharp, and probably erosional. The marl sediment of the Kirtlington Mammal Bed contains subangular pebbles of oolitic limestones, comminuted shell debris, individual ooliths, and rare silica sand grains, all of which suggested to Freeman (1979) a palaeoenvironmental interpretation as a temporary freshwater pool that received periodic influxes of poorly sorted sediment derived from local erosion of earlier Mid Jurassic limestones. The limestone debris is particularly concentrated toward the base of the deposit and was almost certainly derived from the underlying bed. The ostracods, charophytes, and fishes recovered from the horizon are thought to have lived in the pool, and the plants, amphibians, reptiles, and mammals presumably lived nearby (Freeman,

1979). No further palaeoenvironmental work has been attempted for the Mammal Bed assemblage.

11.5.4. Kirtlington Mammal Bed assemblage

The fish fauna recovered from the Mammal Bed is unremarkable in its content, although it is much more diverse than that recovered from the Hornsleasow samples; it contains the typical hybodont shark and halecostomid bony fish components found in most Great Oolite Group assemblages. The microvertebrate fauna recovered indicates the presence of neoselachian teeth (Evans and Milner, 1994), several of which may be attributable to the primitive ray *Spathobatis*, a genus present throughout the British marine Bathonian (Ward, pers. comm., 1993; Metcalf and Underwood, in prep.). The bony fish component consists of disassociated teeth and scales of various holosteans, and these constitute the most abundant fossils within both meso- and microfossil fractions. The commonest of these is the pycnodont *Eomesodon bucklandi*, although *Microdon* and at least one semionotid are also present. Although some specimens are assigned to specific taxa, the vast majority of material is generically indeterminate. Evans and Milner (1994) also record the presence of possible amioid teeth in the acid residues, which Freeman (1976) referred to the Jurassic genus, *Caturus*. As with other Bathonian marine fish assemblages (section 11.7), there is some indication of a prevalence for adaptation to a diet of hard-shelled prey items such as the abundant 'flat clam' bivalves, as well as other molluscs and crustaceans of that time (Ward, 1990) (e.g. the hybodonts *Asteracanthus*, and *Lissodus*, the "batoid" neoselachian, and the semionotids and pycnodonts).

The amphibians include the disassociated and fragmentary remains of a frog referable to the family Discoglossidae (*Eodiscoglossus oxoniensis*), and five species of salamander (*Albanerpeton, Marmorerpeton kermacki, M. freemani*, and two unnamed forms). *Eodiscoglossus oxoniensis* is also present (although much more rare) in the Hornsleasow assemblage (Chapter 9). The record of *Albanerpeton* is one of the oldest of this enigmatic family, the oldest being from the Bajocian of Aveyron, France (Evans and Milner 1994). An albanerpetontid is also known from the microvertebrate accumulation discovered at Hornsleasow. *Marmorerpeton kermacki* and *M. freemani* are the earliest known salamanders (i.e. true Caudata; Evans *et al.* 1988), and are also present in the Hornsleasow assemblage. The remaining amphibian taxa, Kirtlington Salamanders A and B are yet to be described, but are also found at Hornsleasow.

As at Hornsleasow, aquatic turtles are represented in the Kirtlington Mammal Bed by many fragmentary specimens of carapace and postcranial material. The assemblage also includes more diagnostic skull fragments which are thought to derive from a pleurosternid form (Evans and Milner, 1994). The aquatic choristodere, *Cteniogenys oxoniensis*, which is also present at Hornsleasow, was originally described from fragmentary material in the Kirtlington collections at UCL (Evans, 1989, 1990,

1991). The available skull and postcranial material is much better preserved than that from Hornsleasow, however, and indicate that the Kirtlington form is represented by animals of more than one age class, the largest specimens being well ossified and can be assumed to be the adults. The abundant crocodile remains are nearly all shed tooth crowns of the small ?freshwater goniopholid rather than of the marine teleosaurs. Their abundance in the Kirtlington fauna (Fig. 11.1a) suggests an extremely similar palaeoecological situation to that developed at Hornsleasow.

Terrestrial components are also recovered in the microvertebrate accumulation, and include a variety of lepidosauromorphs (Evans and Milner, 1994). As well as the enigmatic *Marmoretta oxoniensis* which was described from Kirtlington and features in the Hornsleasow fauna, true lepidosaurs are represented by some sphenodontids (Evans, 1992) and squamates (two scincomorph lizards, an anguimorph and a possible gekkotan), which are not found in the older assemblage (Chapter 9). *Marmoretta* is apparently a common component of the Kirtlington fauna (Evans, 1991) and was a small, probably insectivorous form. Teeth of pterosaurs and dinosaurs are relatively common in the assemblage and appear to be of the same form and variety as those from Hornsleasow (Benton and Spencer, 1995). The teeth of the tritylodont mammal-like reptile, *Stereognathus*, are also recovered from the Mammal Bed. These represent the youngest known tritylodonts, and the last surviving mammal-like reptiles in Britain, superseded by a Chinese form from the Mid-Late Jurassic (Benton and Spencer, 1995). *Stereognathus* teeth are by no means as common at Kirtlington than they are in the Hornsleasow clays. The assemblage also includes five mammal taxa which were all presumably terrestrial in habit. Their relative abundance in the accumulation is not indicated by Freeman (1979) or Evans and Milner (1991, 1994), although they are by no means as rare as mammal teeth are in the Hornsleasow concentration (Chapter 9).

Invertebrates are in the form of disarticulated and poorly preserved shell debris (crinoid, echinoid, brachiopod, bryozoa, bivalve etc.) derived from the underlying limestones or contemporary storm-washover. Well preserved oyster shells, gastropods, foraminifera and ostracods are also reported, and suggest low or variable salinity's (S. Evans pers. comm.). For instance, the gastropods seem to be generally assigned to various low salinity taxa, such as *Valvata*, *Viviparus* and *Planorbis*, although cerithid and nerinid taxa were also recovered.

Plant remains are found throughout the Mammal Bed and occur in the form of lignitic and limonitic traces, charophyte gyrogonites and megaspores (Evans and Milner, 1994). The palynomorph assemblage recovered from Kirtlington includes similar pteridophyte and gymnosperm taxa as those in the Hornsleasow clays, but with a far greater proportion of fern spores (E. Freeman, pers. comm.). This suggests that Kirtlington was probably more distant from drier or upland areas, occupying a lower coastal region. As at Hornsleasow, the freshwater alga, *Botryococcus braunii*, has been

recovered from the horizon, suggesting that the accumulation formed in clear standing water, with little current or wave action.

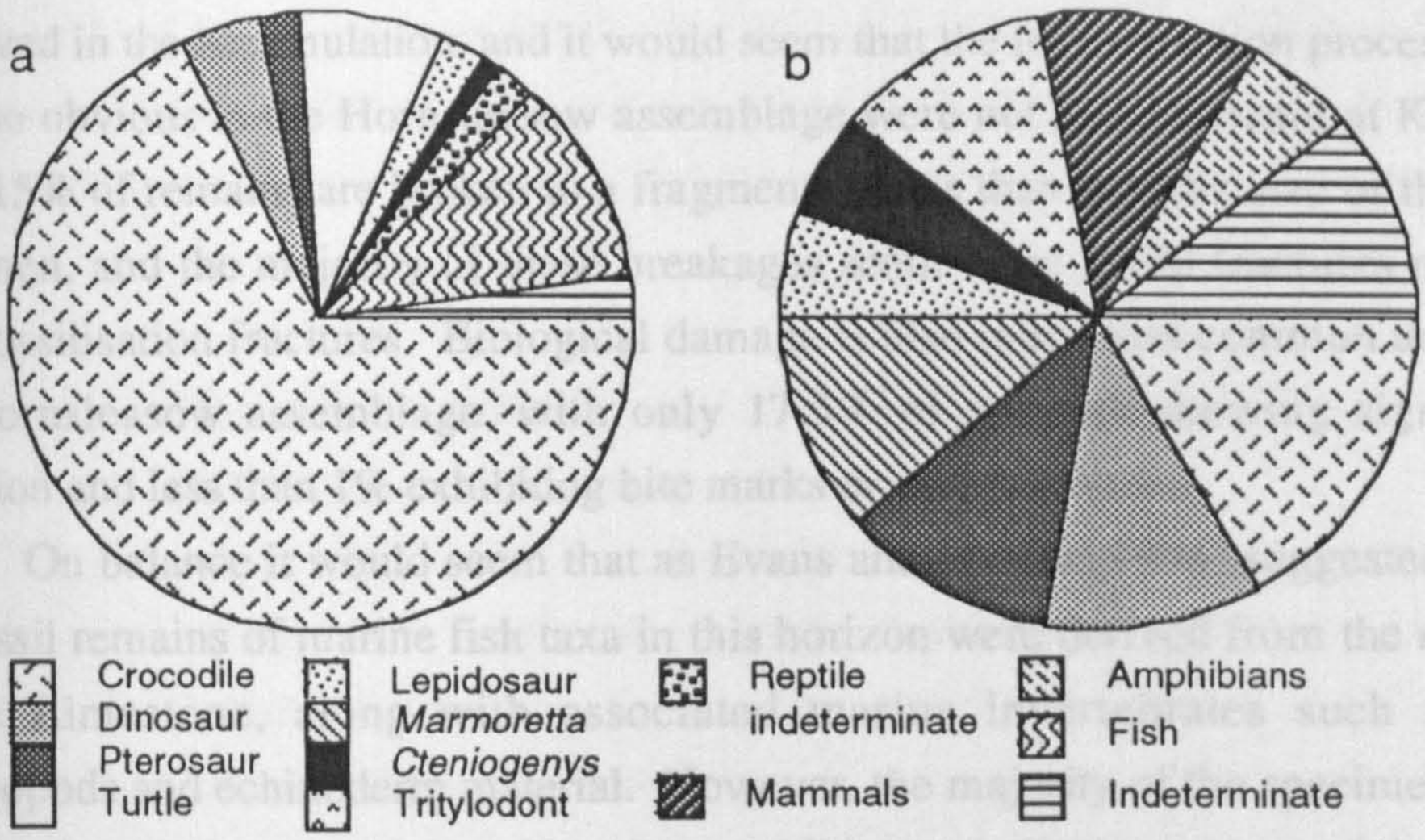
11.5.5. Taphonomy and Palaeoecological conclusions

Evans and Milner (1994) suggest that the vertebrate fauna of the Kirtlington Mammal Bed, with its fishes, amphibians and aquatic reptiles (choristoderes, crocodilians, and turtles), agrees well with a freshwater lagoonal scenario (Table 11.1), although they did not include a taphonomic appraisal of the accumulation. They did state that the terrestrial elements, including the albanerpetonid amphibian, lizards, archosaurs and mammals, are much rarer (Benton & Spencer, 1995) and suggested that these components may have been transported into the lagoon downstream from inland regions.

Evans and Milner provided a list of taxa, but have not stated which forms are rare or common in the assemblage. My studies of the Kirtlington material was carried out at UCL in 1992, when I recorded the preservational features of 90 vertebrate specimens. These specimens had been sampled during a pilot study at Aberystwyth University looking for ostracod fossils (S. Evans pers. comm.). The sample ('3A') came from the main channel-like body of the Kirtlington Mammal Bed, and contained abundant invertebrate material including 128 small gastropods, of which well over half (N = 65) belonged to low-salinity genera. Other invertebrate debris included crinoid ossicles, echinoid spines and fragments of test, nerinid, trochid and cerithid gastropods and *Favreina* sp. (crustacean faecal pellets) (S. Evans pers. comm.). None of this material was retained at UCL and so I have not had the chance to examine it myself. The vertebrate material had not been sorted taxonomically (as the UCL collections had), and so, was more taphonomically comparable to that derived from Hornsleasow.

The results of the study are shown in Table 11.1 and Figure 11.1 (a,c,e,g). Although certain components of the Kirtlington fauna differ from Hornsleasow (e.g. there are more definite marine taxa in the younger deposit : Table 11.1), the assemblages are quite similar in composition. As with the Hornsleasow assemblage, the unsorted Mammal Bed sample showed an abundance of goniopholid crocodile remains (mainly teeth) and a dominance of reptilian forms over all over taxa (Fig. 11.1a,c). As at Hornsleasow, most of the vertebrate material recovered from sampling of the Mammal Bed, occurs as isolated elements (Fig. 11.1c), and shed teeth made up over three quarters of the identified remains. The weathering profile (Fig. 11.1e) shows that weathering was not a significant process in the taphonomic history of the Kirtlington assemblage, although as with the Hornsleasow microvertebrate specimens difficulties were encountered in determining actual weathering stages for such tiny remains. The abrasion profile is highly variable in form (Fig. 11.1g), and some specimens show fairly advanced stages of abrasion and transport. However, many delicate elements are

Figure 11.1. Palaeoecological and taphonomic studies in the Forest Marble Formation at Kirtlington (a,c,e,g) and Dorset (b,d,f,h). Pie-charts show distribution of taxa (a,b) and skeletal elements (c,d) in the studied samples. Histograms show weathering profiles (e,f) and abrasion (g,h) for the specimens. Total number of studied specimens for Kirtlington N = 90, and for Swyre/Watton Cliff N = 78.


a
b
Crocodile
Dinosaur
Pterosaur
Turtle
Lepidosaur
Marmoretta
Cteniogenys
Tritylodont
Reptile indeterminate
Mammals
Amphibians
Fish
Indeterminate
c
d
Teeth
Jaws
Vertebrae
Girdle
Limbs
Indet. bone
Carapace
Fish scales
e
f
Percentages
40
20
10
0
'0'
'0(1)'
'1'
'1(2)'
'2'
'2(3)'
'3'
Weathering Stages
50
g
h
30
Abrasion Indices

preserved in the accumulation, and it would seem that the fragmentation processes which were so obvious in the Hornsleasow assemblage were not as significant at Kirtlington. Only 15% of remains are broken to a fragment of less than half the size of the original specimen, and the majority of these breakages seem to be spiral fractures rather than post-fossilisation fractures. Biological damage is also much less common than seen in the Hornsleasow assemblage, with only 17.5% of remains showing signs of acid digestion and less than 1% exhibiting bite marks or trample marks.

On balance it would seem that as Evans and Milner (1994) suggested, some of the fossil remains of marine fish taxa in this horizon were derived from the underlying White Limestone, along with associated marine invertebrates such as corals, brachiopods and echinoderm material. However, the majority of the specimens which I examined and also material such as the amphibian remains have most of their skeletal elements preserved and these are possibly those which have been least transported. This supports the palaeoenvironmental interpretation given by Evans and Milner (1994) of a freshwater lagoonal setting with marginal marine influences, an indigenous fauna of aquatic and semi-aquatic vertebrate forms, and a steady supply (?by rivers) of terrestrial skeletons and bone material. Coprocoenosis as an accumulating factor for the components of the assemblage seems to have been much less than at Hornsleasow.

11.6. Mixed terrestrial and marine faunas: Watton Cliff and Swyre

The Forest Marble (Upper Bathonian) at Watton Cliff (also known as Ware or West Cliff), West Bay, Dorset (SY 451908 - 453907), contains channel deposits developed within a thick shelly horizon which have yielded abundant microvertebrate material including a reworked terrestrial component (Evans, 1992; Evans & Milner, 1994). The site was excavated in the late 1970's by a team from UCL (Kermack, 1988; Evans, 1992). The fauna consists of selachians, including micro-shark remains, bony fish and importantly, amphibian material comparable to that from the Forest Marble at Kirtlington Cement Works (Evans, 1992; Evans & Milner, 1994; D. Ward, pers. comm., 1993). The site has also yielded abundant small reptilian remains (Evans, 1992), along with much rarer tritylodont mammal-like reptile and mammalian teeth (Freeman, 1976; Ensom, 1977; Kermack, 1988; Kermack *et al.*, 1988). Watton Cliff is an SSSI for both fossil reptiles and fishes (Benton and Spencer, 1995; Dineley and Metcalf, in prep.). A similar fauna has been recovered from the same beds at a cliff-top exposure at nearby Swyre Head (SY 525868).

11.6.2. Geological Setting

The cliff section at Watton Cliff has been described by H.B. Woodward (*in* Strahan, 1898, p.8) and by Torrens (1969, p.37-38). The succession here is the most complete sequence of Forest Marble in west Dorset and additional information on the geology has

been provided by Melville & Freshney (1982, p.26-27), House (1989, p.57-59), and Holloway (1985, p.260). The section given here is based on the work of H.B. Woodward with modern terminology and measurements applied were appropriate:

(Combrash)

Forest Marble Formation

10. Flaggy blue limestone, showing ripple-marks, and clays or shales, with "race"; the limestone preponderating:	3.04m
9. Clays with "race", shaley limestone, thin shelly limestone and thin leaves of sandy limestone, ferruginous in places; the clay preponderating:	6.10m
8. [= 'calcirudite' of Holloway = ?*Digona* Bed of Torrens = 'Mammal Bed' of Freeman] False-bedded shelly limestones, sandy and oolitic in places, with irregular clay-seams, many ochreous galls, lignite; and with the bivalves *Camptonectes, Plagiostoma, Praeexogyra* and fragments of the crinoid *Apiocrinus*:	3.00-4.60m
7. Grey clay (impersistent):	0-0.90m
6. Hard, white or grey marl, with thin seams of bluish shelly limestone:	0.15m
5. Blue flaggy argillaceous limestones, and blue and yellow clays, with thin layers of calcareous grit:	9.15m
4. [= *Boueti* Bed] Hard, sandy marl stained reddish-brown; brachiopods '*Rhynchonella*-bed, with *Chamlys vagans, Goniorhynchia boueti, Avonothyris langtonensis, Ornithella digona,* and crinoid (*Apiocrinus*) ossicles and serpulids:	0.36m
Fuller's Earth	seen 28.00m.

The section of Forest Marble Formation consists of three lithological units of roughly equal thicknesses (Arkell, 1947). The lower unit consists of a thick greeny-brown marly clay interbedded with thin impersistent shelly limestones, silts and sandstone bands (Holloway, 1982). Similarly, the upper horizon is dominated by laminated marly clays interbedded with very fine sandstones and silts. These two units are separated by 3-5m of coarsely bioclastic limestone (bed 8 of Woodward's log), known as the 'calcirudite' bed (Holloway, 1982) and thought to be laterally equivalent to the brachiopod-rich bioclastic *Digona* Bed of the Weymouth region (Torrens, 1969). At Watton Cliff there is another rich brachiopod-bearing, shell fragmental band, known as the *Boueti* Bed (bed 4 of Woodward's log; Melville & Freshney, 1982), which provides

a convenient base for the formation as it is laterally persistent and can be traced northwards into Somerset and the southern Mendips, enabling lithological correlation of the formation with the succession in the southern Cotswolds to be made (Arkell, 1947; Cope *et al.*, 1980).

Detailed biostratigraphic correlation of the Forest Marble sequence in west Dorset with the standard Upper Bathonian zonation is not easy as no diagnostic ammonites have been recorded in these beds. Correlation can, however be demonstrated based, upon the prevalence of the lithostratigraphic marker beds, the *Boueti* and *Digona* Beds, and their characteristic faunas, as ammonites occur in the east Dorset succession and have been recovered from the two brachiopod-rich units. If Torrens' (1969, p.37-38) assertion that the microvertebrate-bearing calcirudites (bed 8) are the lateral equivalent of the *Digona* Bed, then the appearence of the subzonal ammonite *Clydoniceras hollandi* within this unit suggests a precise age for the Watton Cliff assemblage.

11.6.3. Watton Cliff Mammal Bed

The microvertebrates were recovered from the thick calcirudite horizon, known since as the 'Watton Cliff Mammal Bed', in the middle of the section (bed 8) by bulk sampling (Freeman, 1976; Kermack, 1988; Kermack *et al.*, 1987). This unit consists of individually impersistent sheets and lenses of planar or cross-bedded shell-fragmental and oolitic limestones, interbedded with thin marl drapes. The invertebrate macrofauna of the calcirudite units is largely made up of broken valves of *Praeexogyra hebridica* and whole large pectinids. Rarer elements of the assemblage include rhynchonellids and moulds of the bivalves *Gervillia* and *Isognomon* (Arkell, 1947). Much of the biodebris is reported to be abraded, bored and encrusted by serpulids (Holloway, 1982). Carbonised plant matter is frequently observed within this unit and occurs as fine disseminated fragments and as large log material. The vertebrate-bearing channel-like deposits are thought to represent storm breaches through an offshore shell bank complex (Holloway, 1982). The surfaces of individual beds may be ripple laminated and some show small parallel scours, whilst their bases are sharp, suggesting deposition in shallow agitated waters. Elsewhere beds are truncated by channel-like bodies of marl or shelly limestone, which frequently contain rolled and bored homiolithic pebbles of the marly limestone and less commonly large spherical clay intraclasts interpreted as mud balls, similar to those found within tidal channels in the Florida Keys region (Holloway, 1982). Holloway (1982) considered the shell debris forming the calcirudite shoals to be derived from the underlying Forest Marble and that the only organisms living within the subtidal shoals and channels are the boring bryozoans and encrusting organisms. Terrestrial material such as the remains of the small tetrapods and the large pieces of wood were deposited within the cross-cutting channels during the waning stages of storms, and suggest a close proximity to land (?Cornubia). The sedimentary and

taphonomic palaeoenvironment have not been studied for the Swyre vertebrate-bearing horizon, but Evans and Milner (1994) suggested that it is the same as that proposed by Holloway (1982) for Watton Cliff.

11.6.3. Mammal Bed fauna

The vertebrate remains at Watton Cliff were recovered by bulk sieving and acid preparation techniques upon the shelly limestones and marls known as the 'Mammal Bed' (Freeman, 1976; Kermack, 1988; Evans 1992; D. Ward pers. comm.). Much of the material is now housed in the UCL collections (Evans, 1992, and pers. comm.). The microvertebrate fauna recovered from both Swyre Head and Watton Cliff has two well-defined components: the first consisting of well-preserved and clearly marine fish remains; and the second being comprised of derived terrestrial tetrapods (Table 11.1).

(1) *The indigenous marine fauna.* The fish fauna recorded from Watton Cliff is largely made up of microshark material such as teeth and dermal denticles, associated with a small holostean bony fish component of teeth and scales (D. Ward, pers. comm.). The ubiquitous Jurassic hybodont genera are present at both localities and include material from *Hybodus*, *Polyacrodus*, *Asteracanthus* and *Lissodus*, similar to forms recovered from the *fimbriatus-waltoni* Beds at Kirtlington There may also be more than one species of *Hybodus* and *Lissodus* in the fauna, but these have yet to be fully described (D. Ward, pers. comm.).

Neoselachians are also a common component of the microshark fauna recovered from the calcirudites at Watton Cliff (D. Ward, pers. comm.). These include the primitive ray *Spathobatis* and the ?ancestral squalid shark *Protospinax*, both forms adapted to a specialised benthic mode of life as their crushing-type dentition indicates (Cappetta, 1987). Galeomorph sharks are well represented in the Forest Marble fauna of Watton Cliff. A species of the extant heterodontid genus, *Heterodontus*, has been recovered from the acid residues (Ward, pers. comm. 1995) and represents one of the earliest occurrences of the genus and indeed of the Heterodontidae family (Cappetta, 1987). Although undifferentiated heterodontid teeth and denticles have also been described from the Forest Marble of Cirencester, Gloucestershire (Young, 1984). Modern *Heterodontus* is a small benthic shark, which lives in shallow, warm waters and possesses a differentiated clutching-grinding dentition which has undergone little modification from the earliest occurrence of the family to the present (Cappetta, 1987). An orectolobid shark (Order Orectolobiformes) is also present in the acid residues, but generic diagnosis is uncertain (D. Ward, pers. comm.). Orectolobids are a common neoselachian component of many Bathonian faunas, and have been recovered from the Middle and Upper Bathonian of the Cotswolds (Young, 1984; Metcalf and Underwood, in prep.; section 11.7). The Lamniformes are represented in the Watton Cliff assemblage by *Palaeocarcharias*, a genus described from only three partial specimens

from the Lithographic Stone (Upper Kimmeridgian) of Eichstätt, Germany (de Beaumont, 1960). The recognition of its teeth in the Watton Cliff assemblage is the earliest occurrence of the genus and possibly the order (Cappetta, 1987). Scyliorhinid teeth (Order Carcharhiniformes) have also been recognised in earlier Bathonian deposits at Huntsman's Quarries, Gloucestershire (Metcalf and Underwood, in prep.; section 11.7) and in Africa (Cappetta, 1993, Cappetta *et al.*, 1994) and teeth of the scyliorhinid *Scyliorhinus* are recorded from the Forest Marble at Watton. Neoselachians are not reported from the Swyre horizon (Evans and Milner, 1994).

(2) *The derived component.* A variety of freshwater aquatic and terrestrial forms similar to those described from Kirtlington and Hornsleasow are found in the calcirudites of Watton Cliff and Swyre Head. The fauna includes amphibians, reptiles and mammals, and all appear to have been derived (along with the plant debris) from nearby land.

The amphibians recorded from the Forest Marble at Watton Cliff have yet to be fully described (Evans, 1992; Evans & Milner, 1994), but the composition of the fauna seems to be typical of late Bathonian tetrapod assemblages (cf. Evans *et al.*, 1988, 1990). The fauna includes all the major components of the Kirtlington Mammal Bed (section 11.5) except for the caudate, 'Kirtlington Salamander B' (Evans, 1992).

Similarly, the Kirtlington aquatic reptile forms are also recorded in the assemblage (Table 11.1), including turtle, choristodere and goniopholid crocodile material (Evans and Milner, 1994). Unusually for a Bathonian marine assemblage, no teleosaurid crocodile teeth or bones were reported by Evans and Milner (1994).

The terrestrial component of the derived assemblage comprises indeterminate pterosaur remains, the teeth of several dinosaur taxa, and the tritylodont *Stereognathus*. Several mammalian taxa have been recovered from the sampling of the calcirudites at Watton Cliff, but only indeterminate mammalian teeth have been found at Swyre. However, whether this discrepancy is due to palaeoecological, taphonomic or sampling procedures is not discussed by Evans and Milner (1994).

11.6.4. Taphonomy and palaeoecological conclusions

Only the derived component of the Dorset microvertebrate accumulation was housed at UCL, and therefore I could only make a study of this material (Fig. 11.1b,d,f,h). The taxonomic pie-chart (Fig. 11.1b) shows an unrealistic spread of distribution of taxa in this sample, and this is largely due to sorting by the investigators at UCL during sampling procedures. Therefore no palaeoecological information can be derived from this distribution. This is also reflected in the second pie-chart (Fig. 11.1d), which shows a much more even distribution of skeletal remains than either the Kirtlington or Hornsleasow assemblage. However, sorting by transportation could also have been responsible for the preservation of so many easily transported skeletal elements, such as

vertebrae and girdle fragments. Although more streamlined elements, ribs, jaws and limb bones are also equally well represented in the sample.

The weathering profile is somewhat variable (Fig. 11.1f), although generally indicating low stages of subaerial erosion. Similarly, no specimens which I studied showed any evidence of bite, gnaw or trample marks, although 24% did show evidence of stripping by acid digestion. According to David Ward (pers. comm.) the indigenous fish material comprises isolated scales, teeth and bones, all of which are extremely well preserved, exhibiting no signs of wear. However, the tetrapod material which I studied is extremely fragmentary and generally waterworn (Fig. 11.1h), indicating considerable transport into the high-energy offshore depositional environment. All spiral fractures show rounding. Transverse fracturing is rare, although post-fossilisation mineralisation (iron oxides) is seen encrusting the surfaces of bones and sometimes in fine cracks in the surface.

11.7. Shallow marine palaeoenvironment: Eyford

The newly defined Charlbury Formation (*progracilis* Zone: Middle Bathonian) is largely composed of shallow open marine carbonates (Boneham & Wyatt, 1993). However, in the Eyford area, west of the Cotswolds town of Stow-in-the-Wold, the Charlbury Formation is composed of flaggy, arenaceous limestones, known as the Eyford Member (Sellwood & McKerrow, 1974). These have been quarried in this area for tilestone since the nineteenth century, and were once thought to be a lateral continuation of the 'Stonesfield Slates' of Stonesfield, Oxfordshire (SP 387171), which have now been re-assigned to the overlying Taynton Limestone Formation (*progracilis* Zone: Boneham & Wyatt, 1993).

The Eyford Member lithology (formerly known as the 'Cotswolds Slates', Richardson, 1929) contains an abundant and well-preserved vertebrate fauna, as well as plant fragments and invertebrates. This unit and its biota have received far less attention than the more famous 'Stonesfield Slates'. This is partly because the Oxfordshire site yielded *Megalosaurus bucklandi,* the first dinosaur to be named (Buckland, 1824), and the original Jurassic mammal and therapsid discoveries (Buckland, 1828, Phillips, 1871, Woodward, 1894). The 'Cotswolds Slates' have not yielded any therapsid or mammalian remains. Secondly, the latter was close to Oxford University, which was the centre of the British academic world at the time the discoveries were made (Savage, 1963). The Eyford Member localities are centred around Eyford Hill (SP132252) and Sevenhampton Common (SP0122) in Gloucestershire. However, large collections of Eyford material (often labelled as "Stonesfield Slate, Eyford") were made by gentlemen scholars and amateurs in the nineteenth century, and one of these, the extensive Witts collection, housed at Gloucester City Museum, formed the basis for the only systematic palaeontological work on the 'Cotswolds Slate' (Savage, 1963).

Despite its historical importance and rich fauna, no modern palaeoenvironmental review of the 'Stonesfield Slate'-type facies has been completed. Although the tilestones are no longer exposed at Stonesfield itself, the Eyford Member is readily accessible in a number of working and partially infilled quarries in the study area. In 1992-3, I undertook an analysis of the sedimentology and composition of the faunas of the Eyford Member and those of the overlying Taynton Limestone Formation marine oolites in collaboration with Dr. C. Underwood of Liverpool University. Our studies are to be published in a paper (Metcalf and Underwood, in prep.) and allow the construction of a palaeoenvironmental and taphonomic model for these fossiliferous facies.

11.7.1. Geological setting

The study area was taken to encapsulate the main outcrop area of the laminated tilestone facies within the Charlbury Formation (Eyford Member). These rocks are restricted geographically by overstep of the Taynton Limestone Formation to the north and east, by erosion and partial gradation into the Sevenhampton Marl Member to the west and by gradation into the more open marine micritic and oolitic limestones of the Farmington Member and Charlbury Formation *ss.* to the south. The Eyford Member outcrops in a series of irregular outliers spread over about 70 km^2 of the northernmost Cotswold hills, and was formerly worked extensively in the Eyford area for roofing slates. However, the Eyford Member is now only actively worked in a small number of quarries, and due to the relative rarity of vertebrate fossils, it was only practical to collect faunas from actively working quarries. The majority of our work was concentrated on the two largest of these, Huntsman's Quarry (SP1225) and Brockhill Quarry (SP133236), both near the village of Naunton, some 15km east of Cheltenham Spa. The Charlbury Formation *ss.* is worked elsewhere in the Cotswolds region (Boneham & Wyatt, 1993), but vertebrate fossils are not abundant within this facies.

Despite a far larger outcrop area of the Taynton Limestone Formation, extending across most of the northern Cotswolds, the biooosparites of this formation are less well exposed than the Eyford Member due to their poor quality as a building stone. Within the study area they are only being actively worked at Huntsman's Quarry. None of the previous collections of fossils were made from this lithology, and it is possible that the historic quarrying activities at Eyford and Sevenhampton were only in the tilestone. There could also be a factor of collector bias, as the fossils from the Taynton Formation are often broken and heavily abraded.

The Charlbury and Taynton formations occur low down within this sequence and are typical in sedimentary character of the succession as a whole, being composed on the whole of shallow marine carbonate facies. The vertebrate-bearing laminated tilestones of the Eyford Member are a localised facies variant of the Charlbury Formation in the northern Cotswolds. They are older than the more famous tilestones of the

'Stonesfield Slates' of Oxfordshire, which are concentrated in the overlying Taynton Formation (Boneham & Wyatt, 1993). There are however, thin tilestone partings within the Taynton Formation sections at Huntsman's Quarry. Both the Charlbury Formation and the Taynton Formation lie within the lower Middle Bathonian, in the *Procerites progracilis* ammonite Zone.

The area of Eyford Member deposition was upon the Cotswold-Weald carbonate shelf, situated at the western transitional boundary from the shelf into the Severn basin, passing into the quiet marine-clays of the Fullers Earth in the south-west Cotswolds and around the Bath area (Arkell & Donovan, 1952). Deposition on the carbonate shelf, in the Cotswolds, grades laterally into the 'Midland shallows' in north-east Oxfordshire and Northamptonshire, whose relatively thin deposits of marine, lagoonal and coastal-marsh sediments (Rutland Formation, formally the Upper Estuarine Series, Hallam, 1992) are bounded to the east by the low-lying Anglo-Brabant landmass. The shoreline of the Pennine-Welsh landmass to the north and north-west of the area may have been close to the study area, and provided the source of the terrigenous material within the Eyford Member (see discussion below, Cope *et al.*, 1992, Metcalf, 1994).

11.7.2. Sedimentary facies definitions

The Charlbury Formation is represented in the studied area by the sandy tilestones and associated limestone facies, and is known as the Eyford Member (Sellwood & McKerrow, 1974). To the east of the area (and east of the Moreton Swell structure) the Charlbury Formation is represented by shell fragmental limestones and subsidiary marls. In the type area at Charlbury (SP 370200) the shelly limestones reach a thickness of 3-4m, while to the south-west near Cirencester the Formation reaches 8m (Boneham & Wyatt, 1993; R.J. Wyatt, pers. comm.). The base of the Eyford Member in the Eyford area occurs within a continuous sequence over the Lower Fullers Earth Clay (Fig. 11.2). Although this boundary is nowhere well exposed, from a temporary section at Brockhill Quarry, it can be seen to be fairly gradational in character (Fig. 11.2). Directly overlying the Fullers Earth Clay are silty and microoolitic limestones, with thin shelly intercalations, of the typical 'tilestone' facies. For the first 1.5 metres these are interbedded with thin partings of macrofossil-poor grey-green clay. These clay partings are similar to the Fullers Earth and microfossil sieving of both units have revealed the same sparse marine microfauna dominated by broken oysters, echinoid and asteroid ossicles, foraminifera, and fish debris. The Fullers Earth is replaced to the north and east by cross bedded oolites of the diachronous upper part of the Chipping Norton Limestone Formation, and the overlying clays and marls of the Sharps Hill Formation. This change is relatively coincident with the overstep of the Taynton Limestone and removal of the Charlbury Formation (Boneham & Wyatt, 1993).

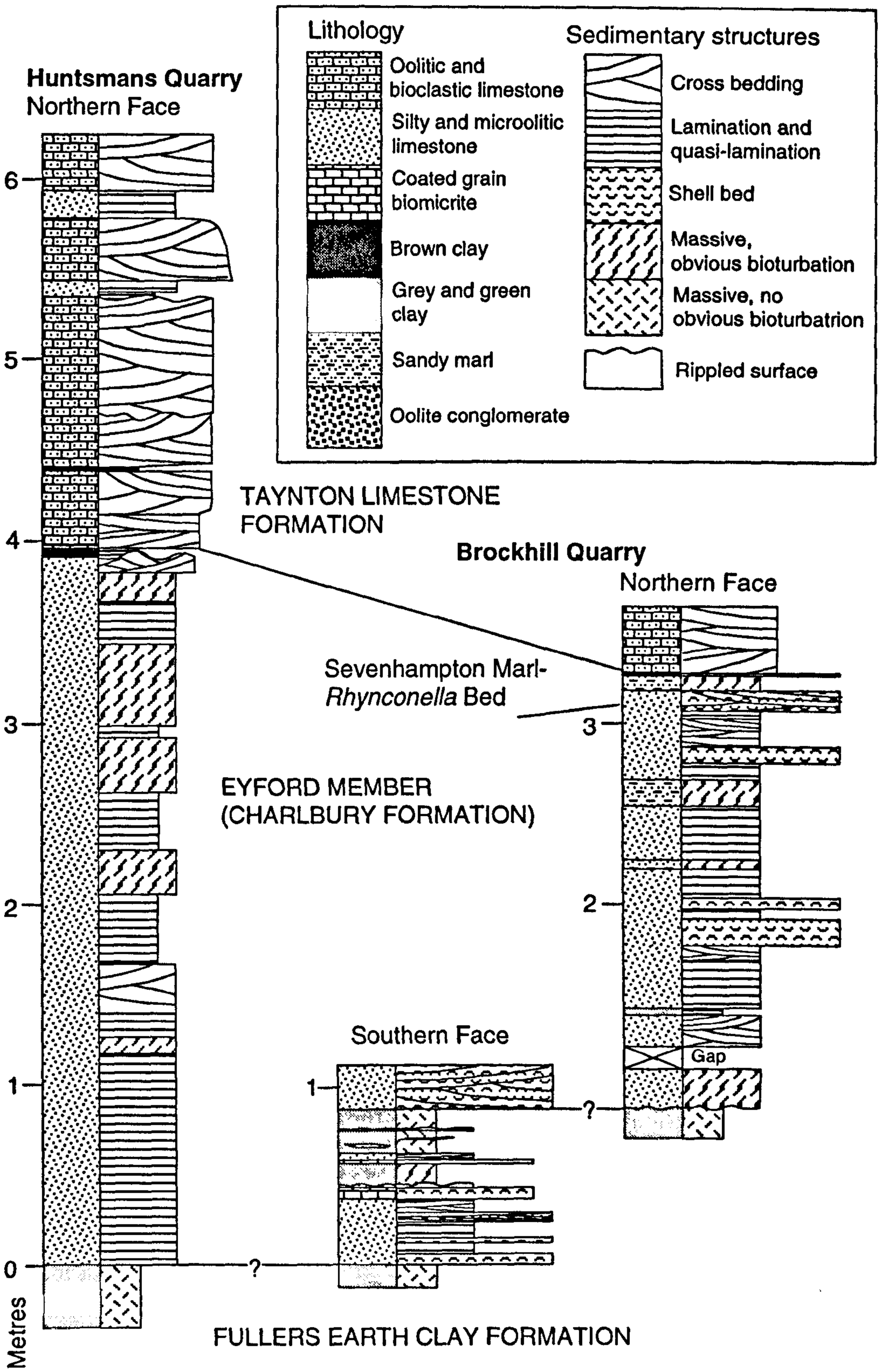


Figure 11.2. Graphic logs showing the succession through the Fullers Earth Clay Formation, the Charlbury Formation and the Taynton Formation in the Eyford area.

There are three lithofacies differentiated in the Charlbury Formation at Eyford (Metcalf and Underwood, in prep.):

1. *The 'tilestone'-facies.* The Eyford Member is extremely laterally variable. This is due partly to lateral facies changes but more significantly to considerable penecontemporaneous erosion. The tilestone facies is the most diagnostic lithology, and makes up between 50-60% of the Formation (Boneham & Wyatt, 1993; Fig. 11.2). They are lithologically fine-grained, calcareous sandstones and siltstones. It is their high siliceous content which makes them such excellent roofing tiles and building stones. They are extremely hard, and are pale yellowish-brown in colour, weathering to a warm greyish-orange on exposure. A very hard, superior building-stone variety of the tilestones, known as "Cotswolds Blue" due to its unweathered bluish-grey hue, is also quarried at Brockhill. In petrographic section the sand and silt grains are seen to be superficially micro-oolitic (Metcalf and Underwood, in prep.). Occasional, discontinuous thin intercalations of a dark-green clay is seen coating the upper surfaces of individual tilestone units. These clay lenses contain a brackish to normal marine microfauna and are frequently well-bioturbated. Within the normal tilestone facies shell fragments and reworked pebbles of oolite and marl, make up less than 5%, although locally there are impersistent shelly and oolitic bands (Sellwood & McKerrow, 1974). Plant fragments preserved as iron-stains are abundant within the Eyford Member. Carbonised debris disseminated throughout the tilestone and basal clay beds is also common. Thinly parted, otherwise unfossiliferous plant-rich tilestones occur toward the base of the Eyford Member sections in Brockhill Quarry (Fig. 11.2), the iron-stained plant material in these concentrations is frequently current aligned (Metcalf and Underwood, in prep.).

The tilestones are extremely fissile, especially when weathered, and this fissility occurs along laminations in the rock. The tilestones appear to be planar laminated, which gives them their flaggy appearence (Sellwood & McKerrow, 1974), but individual laminations are truncated laterally over a distance of a few 100 millimetres (Metcalf and Underwood, in prep.). The laminations make up bed-forms of about 0.1-1 metre thick, which are also discontinuous laterally over a distance of a few 10's of metres. This lateral discontinuity is not obvious in quarry section, and the bed-forms are termed 'quasi-planar' laminated (Arnott, 1993). The base of the beds are often sharp, undulatory, with large scale scouring over several metres. These basal scours contribute to the overall appearence of lateral lensing of bed. Isolated, spatulate flute marks (between 50-100 millimetres in length) were seen upon the base of several tilestone blocks in Huntsman's Quarry, unfortunately palaeocurrent flow direction could not be ascertained as these blocks were not *in situ*..

In lithology, the beds are internally homogeneous and are not usually graded, although shell material, inorganic clasts and charcoal fragments are seen to be

concentrated in the basal scours (Metcalf and Underwood, in prep.). Parting lineation on the surface of individual laminae or beds was not observed and this may be because of the fine grainsize of the micro-oolitic tilestones. The quasi-planar laminated interval usually has an erosional, often ripple-marked upper surface. The ripples range from large symmetrical and asymmetrical wave-formed ripples, with wavelengths of 100 millimetres and sinuous, continuous crests to small scale, poorly developed, discontinuous ripples with wavelengths of about 10 millimetres, resulting from a combined flow regime (Metcalf and Underwood, in prep.). The features of the tilestone beds suggest that they were laid down in extremely shallow, agitated subtidal conditions.

Trace fossils within the tilestone facies are generally common, but of low diversity. Vertical burrows are seen to penetrate the rippled surface into the quasi-planar laminated bed. They include small *Skolithos* (diameter one-two millimetres), and they represent opportunistic trace-makers, colonising the calcareous sand shortly after deposition. They are abundant within the shell-poor laminated tilestone and are least common in the plant-bearing thinly-bedded tilestones seen at the base of the sections. They are nowhere observed to penetrate the intermittent clay lenses.

At Huntsman's, around 50% of the arenaceous limestones are heavily bioturbated, destroying any original lamination or bedding features. These massive beds between 0.1-0.4 metre thick, are often intruded by large *Skolithos*-type trace-makers, which may penetrate the entire thickness of the bioturbated interval and are 20-30 millimetres in diameter. Lithologically, the bioturbated beds are the same as the tilestones, although they tend to be composed of fine sand rather than silt. The section at Brockhill contains only three-four thin bioturbated sandy limestone and marly beds. Although large scale burrows were observed within the beds, the unconsolidated nature of the sediment made elucidation of the trace makers difficult, although vertical burrows appear to be of the *Skolithos*-type.

Other trace fossils are rare, although poorly preserved *Thalassinoides* and *Diplocratarion* have been seen restricted to certain bedding planes in the upper part of the sections at Huntsman's Quarry. Rare *Planolites* are also present and a single specimen of the U-shaped burrow *Rhizocorallium* was recognised in the tilestones at Brockhill Quarry. The trace-fossil *Palaeobullia* commonly covers planar bedding surfaces, but are most abundant upon rippled surfaces. These are thought to represent gastropod grazing trails after restoration of quiet water-conditions. Within the clay horizons are many fine branching burrows filled with micrite. These are sub-horizontal and only penetrate the top 20mm of the clay seams.

2. *Biomicrite-facies.* At Brockhill Quarry the Eyford Member is condensed to around three metres (Fig. 11.2), and contains a number of separate thick, but laterally discontinuous, bands of a very hard, shell-fragmental limestone at various

horizons. These are greyish-orange to dark orange (when weathered) in colour, and are compositionally biomicrites. They contain a substantial amount (15%) of siliceous grains, in the form of silt particles. The biomicrites are generally composed of broken bivalves, echinoids and gastropods (75-85%), reworked small pebbles of oolite, tilestone, marl and clay and disseminated ooliths (15-25%) set within a micritic limestone, and are texturally floatstones (Metcalf and Underwood, in prep.). Most beds are internally graded from a coarse closely packed grainstone or rudstone at the base, into the main floatstone biomicrite which is capped by a layer of shelly tilestone with an upper, rippled surface.

Individual beds of the biomicrites pinch-out laterally, and are in the region of 50-100 millimetres in thickness. Over 70% of the biodebris is broken within had specimens, but whole shells and valves are preserved within the upper floatstone part of beds. Many of these show geopetal structures preserved as sparry, calcitic infills. The shells within the biomicrites are concordant with bedding, as less than 20% are orientated at angles greater than 30° from the bedding plane. The bioclasts within the main part of the biomicrite beds, are 'floating' in the matrix, but they do not show preferred convex-up orientation and therefore exhibit 'disordered stacking' (Kidwell *et al.*, 1986). The grainstone and rudstone, lower portions of the biomicrite beds also show disordered stacking, and in places small bioclasts and inorganic clasts are nested within larger shells (Kidwell *et al.*, 1986)

The biomicrites contain a much wider diversity of macroinvertebrate forms, than the normal tilestone facies. They also contain many more vertebrate remains than the other facies. They appear to lack an, except for some poorly defined escape structures seen in hand specimens. However, this may be a factor of the homogeneity of the biomicrites and there may be unrecognised structures within some beds. All these features suggest that the biomicrites represent shelly channel deposits.

3. *Coated-grain bio-oomicrite -facies.* The lower beds at Brockhill have a much finer grain-size, with less whole shells, and are made up largely of small oyster debris and echinoid spines, and abundant plant remains. They are texturally grainstones and represent attritional deposition in a quiet marine environment. These fine biomicrite beds are often gradational into shelly tilestones and may have a rippled tilestone upper surface (Metcalf and Underwood, in prep.). Those in the lower finer-grained beds are often stacked in the vertical plane, but again exhibit no preferred orientation of valves.

4. Taynton Formation lithofacies. Overlying the Charlbury Formation at Eyford, Richardson (1929) reported the presence of the thin '*Rhynchonella* Bed-Sevenhampton Marl' at Huntsman's Quarry, which he described as a reddish-brown, mottled marl, packed with whole oysters, reworked pebbles, and fragments of wood. The mottling is interpreted as being due to intense bioturbation (Sellwood & McKerrow, 1974). The bed is no longer exposed within the quarry section, although it was located

in rock piles in the western (and oldest) part of the workings and produced two fine large teeth of the mollusc-eating shark *Asteracanthus*. The bed was also found *in situ* within a temporary face in the northern part of Brockhill Quarry (Fig. 11.2). Sellwood & McKerrow (1974) described the bed as having "an erosive base above a trough cross-bedded sandstone[tilestone]" (Sellwood & McKerrow, 1974, p.205) and interpreted that the whole unit was deposited instantaneously as a storm-bed. However, the upper 50 millimetres contain bored tilestone pebbles and mudstone rip-up clasts (Richardson's '*Rhynchonella* Bed', 1929, p.114; pers. obs. at Brockhill, Fig. 11.2) set within a oomicritic limestone. It is likely then that this upper bed is separate and developed upon the erosional upper surface of the 'Sevenhampton Marl' (*sensu* Richardson, 1929). As the basal contact of the Taynton Formation is invariably erosive (Sellwood & McKerrow, 1974, R.J. Wyatt, pers. comm.), the '*Rhynchonella* Bed' is taken by Metcalf and Underwood (in prep.) as the base of the Taynton Formation at this locality (Fig. 11.2).

The overlying Taynton Stone or Freestone (Woodward, 1894) was formalised as the Taynton Limestone Formation by McKerrow & Kennedy (1973). In the type section the Taynton Formation is dominated by cross-bedded bio-oosparites deposited as oolite banks in a mid-shelf environment (Horton *et al.* 1987). The limestones comprise cream to pale yellowish brown and iron-stained (when weathered) flaggy, shelly oolites with thin, impersistent argillaceous and marly bands (Horton *et al*, 1987) and are lithologically variable over the northern Cotswolds. The sandy planar-laminated limestones or tilestone facies of the 'Stonesfield Slates' in Oxfordshire have recently been assigned to several separate tilestone beds within the Taynton Formation and at Huntsman's Quarry there are one or two thin, laterally discontinuous tilestone beds (some with intraformational pebbles developed) in the Taynton limestones.

The sparry bioclastic limestones make up about 80% of the Formation in the Eyford area (toward the east this percentage is reduced to around 50-65%, Horton *et al.*, 1987), and are texturally grainstones and rudstones. Compositionally these limestones are very similar being made up of disseminated oolites (40-50%), bioclasts (around 40%) and few mudstone, marl and oolite pebbles. The finer grained limestones are much more oolitic than the rudstones, and the oolites tend to be superficial grains of fine-sand and bioclasts (Metcalf and Underwood, in prep.). These are also slightly micritic in character, although the limestone is cemented by sparry calcite. The coarse-grained bio-oosparites contain recognisable biodebris, in the form of bivalves, gastropod and echinoderm material. They also contain up to 10% small intraformational micro-oolite and marlstone pebbles, suggesting reworking of the finer fractions (Metcalf and Underwood, in prep.).

The Taynton Formation bioclastic oolites are well bedded, with individual beds varying from about 0.4-1.0 metre in width. The finer oolites generally have slightly

thinner bedding intervals. Internally the beds are often graded, with the basal parts of beds being composed of randomly stacked grainstones and rudstones, consisting of broken shells and pebbles, grading up into finer oolitic-dominant sparry limestones. The shells are closely packed in a vertical plane, and some beds appear to show a 'convex-up' preferred orientation. On a larger scale, beds are cross-stratified, and both trough and herringbone cross-bedding were observed in the sections at Huntsman's.

There are several mudstone and marly intercalations within the Taynton Limestone Formation. The mudstones are lithologically variable and are described by Horton *et al.* (1987, p.82) as "barren, blocky silty mudstone, shelly mudstone or lumachelle, and mudstone with thin silt partings and wisps of shell debris" and these seem to grade into silty laminated beds. At Huntsman's Quarry, Metcalf and Underwood (in prep.) noted several thin, impersistent mudstone beds, which seem to be concentrated in the first half metre of the Formation. They are greenish-grey in colour, and often contain beautifully preserved plant material.

11.7.3. The Eyford biota and biofacies

Over a period of two years Dr. Underwood and myself, in association with various collecting parties from Bristol University made a collection of around 400 vertebrate fossils from the Naunton quarries (Metcalf and Underwood, in prep.). No vertebrate material, however abraded or fragmentary, was discarded unrecorded. Field recordings of material included identification of the fossil, and rock type; orientation on the slab or bedding plane; association with other vertebrate remains, and invertebrates; size measurement (based on long axis length); general preservation, including an idea of how abraded the material is (this is a qualitative process and was therefore ascertained by a single person (me) to keep measurements constant). Only when all these criteria were known was a substandard specimen discarded in the field. However, in all about 85% of the vertebrate remains were collected for preparation and taphonomic study in the laboratory (Metcalf and Underwood, in prep.).

During field sampling a time and motion study was performed to try to quantify the approximate fossil content and diversity of the Eyford Member and the Taynton Formation limestones (Metcalf and Underwood, in prep.). This study focused on the vertebrate remains visible to the naked eye, which are usually remains over two millimetres, and is biased against anything smaller. To test the collecting-bias, samples of each of the limestone facies were prepared in 5% acetic acid solution. This method revealed a more diverse fauna than field collecting alone (23 species versus 12 species from the field study), and produces a more accurate measure of fossil content of the different rock types (Metcalf and Underwood, in prep.). The improved faunal diversity is based upon recovery of micro-sized specimens of under two millimetres such as neoselachian and ray teeth, in terms of this study they shall be termed the

'microvertebrate' content. In the taphonomic study elements 2-20 millimetres in length, such as the many pycnodont teeth are termed 'mesovertebrate-sized' specimens and those above 20 millimetres are termed 'macrovertebrate' remains, these are largely represented by the reptile teeth and bones.

The authors collections were donated to Bristol University Geology museum (BRSUG 25616-26000). Material was also studied from museum collections, in particular the Witts collection, housed in Gloucester City Museum (GLRCM G. 1-722) and Bristol University Museum (BRSUG "Witts collection" various individual numbers). Also studied was the Royce collection in Bristol City Museum (BRSMG; these are on loan from St. Edward's Hall, Stow; Loan number 562 M, Acquisition number 370/1965). Other material has been located in the British Museum (Natural History) (BMNH), Oxford University Museum (OUM) and the Sedgwick Museum (Cambridge) (SMC).

Much of this material is very poorly documented, often being located simply as 'Stonesfield Slate, Eyford', and so it is only rarely possible to identify older material to its particular quarry and all the material *in situ* with associated matrix is from the Eyford Member lithologies, and of this, about 80% is within the tilestone facies. The Witts material was however catalogued and studied by R.J.G Savage (with contributions by D.T. Donovan, 1963, table 1), who gave a vertebrate species diversity list for the Eyford Member tilestone facies of 24. The Royce collection appears to have been made from the same localities, but it cannot be ruled out that some of the material may have come from the Sevenhampton Common quarries to the west.

Although vertebrate fossils are important elements within the Charlbury and Taynton formations, they are never a dominant component being exceeded in both numbers and diversity by invertebrates and occasionally plant remains. There are three distinct biofacies of non-vertebrate remains (Metcalf and Underwood, in prep.), which are roughly constrained by sedimentary facies and the corresponding energy of the depositional environment; these biofacies are:

1. *Low diversity mollusc-biofacies I.* This is dominated by small epifaunal and shallow infaunal molluscs, and is seen in the bulk of the laminated arenaceous tilestone-facies of the Eyford Member and Taynton Formation (Metcalf and Underwood, in prep.).

The ubiquitous Bathonian oyster, *Praeexogyra acuminata* is present within all the biofacies and sedimentary facies (except for the clay laminae). It forms monospecific accumulations within the tilestones facies, but shells are never in sufficient numbers to create oyster banks. The small thin-shelled trigonid, *Vaugonia impressa* is also frequently found in monospecific stringers of disarticulated valves. This species is characteristic of the tilestone facies and is abundant within the Stonesfield Slates *ss.* (Taynton Formation) of Oxfordshire, it was once taken as evidence for the lateral

correlation of the Stonesfield Slates with the Eyford Member tilestone facies (Savage, 1963). *Vaugonia* is most common within the low-diversity mollusc assemblages of the tilestone facies at Huntsman's Quarry, it is rare in the more oyster-rich 'shelly-tilestones' of Brockhill.

Pectinids are also represented within this biofacies, with *Radulopecten* being most frequently recorded. Species of *Camptonectes* and *Plagiostoma subcardiiformis* are also found. Although a range of bivalves are present within the tilestone facies, with 18 species being recorded in the Witts Collection (Savage, 1963), these are extremely rare elements of the biofacies and the only other species to form concentrations are the thin-shelled, epifaunal *Placunopsis fibrosa* and small *Modiolus*.

Stringer accumulations of the small, shallow-infaunal gastropod *Proceritherium (Rhabdocolpus) variabile* occur within this biofacies, although other gastropod species are rather rare (Savage, 1963, table 1). Concentrations of *Proceritherium* have been noted in association with large vertebrate fossils (Metcalf and Underwood, in prep.).

Cephalopod fossils are not common within this biofacies, although the frequently recorded extreme rarity (e.g. Savage, 1963) is probably because of their poor preservation and, hence, poor representation within museum collections, rather than actual impoverishment. Belemnites (apparently all *Belemnopsis fusiformis*) are more abundant than ammonites within this biofacies, but they form a minor part of the invertebrate fauna. The large zonal ammonite *Procerites progracilis* and some nautiloid specimens are also encountered regularly within this biofacies (pers. obs., Savage, 1963).

Within the Witts collection there are four beautifully articulated and several less perfectly preserved specimens of the asteroid *Asteropecten*. These were found within fine-grained storm laminae on Eyford Hill. These are extremely rare in the field, although disarticulated echinoderm material is common within the more shelly tilestone, often concentrated as thin bioclastic laminae within basal scours. The composition of these beds is variable, but generally millericrinid material dominates, associated with ?diademid spines, isocrinid ossicles and fragments of regular echinoid test.

Other invertebrate groups are generally rare within this and the other biofacies, in contrast to other marine Bathonian formations (Metcalf and Underwood, in prep.). Brachiopods are extremely rare in all facies, being represented by rare rhynchonellids and *Epithyris* (pers. obs., Savage, 1963). There is only one specimen of a coral from this biofacies (in the Royce collection), and this is the colonial *Isastraea*. No bryozoa have been recorded in this or the other biofacies, and in contrast to the abundant specimens within the Witts collection (as noted by Savage, 1963), serpulids were not found within the arenaceous tilestones during this study. The tilestones have yielded occasional decapod and cirripede material.

Terrestrial invertebrates are represented by a fauna of insects and beetles within this biofacies, although only within the older museum collections (Savage, 1963, and as noted by M.D. Crane in the BRSMG Geology File Number 54, 1983). Members of several orders are present in the collections, but the assemblage is dominated by large (30-40mm length), well preserved coleopteran elytra. Although it may be in part a failure of collection, the terrestrial arthropod fauna appears to be of a lower diversity than that at Stonesfield (Savage, 1963).

The tilestone facies of Eyford and Sevenhampton have long been well known for their abundance of a rich and well-preserved terrestrial flora (Strickland & Buckland, 1844, Savage, 1963). These remains include a range of seeds, fronds and leaves of cycads, conifers, ginkgoes and ferns (cf. Savage, 1963, table 1). However, these recognisable plant remains are far less abundant than indeterminate woody material, generally in the form of slender branch fragments, which coat bedding planes at some horizons and are frequently current-aligned.

2. *High diversity mollusc-echinoderm-biofacies.* This is recognised within the coarse grained biomicrite facies of the Eyford Member, seen at Brockhill and Salperton Quarries. The biomicrites are rich in infaunal and epifaunal molluscs and echinoderm debris (Metcalf and Underwood, in prep.).

The epifaunal bivalve component of this biofacies is similar to the low diversity mollusc biofacies I, in being dominated by the oyster *Praeexogyra acuminata* and the trigonid *Vaugonia impressa.* Likewise, pectinids are represented by several species, and as with the tilestone-facies *Radulopecten* is the most frequently encountered in field collections, whilst *Plagiostoma* and *Camptonectes* are much more rare. The oyster debris makes up over 40% of all recognisable bioclasts within this rock-type. *Vaugonia* occurs within localised concentrations, in association with thicker-shelled trigonids (cf. *Trigonia* sp.). A similar range of small mytilaceans occur with this biofacies, with *Modiolus* and *Placunopsis* forming localised concentrations. This biofacies also contains a range of bivalves which are rare or absent within the other facies. Particularly obvious are the shallow infaunal forms, such as thick -shelled astartids and large *Girvillella* , and the thin-shelled ?*Thracia* (Metcalf and Underwood, in prep.).

Large gastropod debris is common within the basal scours of storm beds, although much of this is undifferentiated. The small gastropod *Proceritherium* is also observed in the finer-grained portions of the graded beds. Cephalopods are much more common within this biofacies than in the tilestone beds, with the belemnite *Belemnopsis fusiformis*, rare nautiloids and the ammonites *Procerites* sp. and *Oppelia* being recorded during field studies (Metcalf and Underwood, in prep.).

Echinoid debris forms an important component of the biofacies. The composition of the assemblage remains similar to that within the tilestone biofacies, although there is a higher frequency of both isocrinid ossicles and large irregular

echinoid material in the biomicrites. This is probably a reflection of the higher energy deposition of the biomicrite-facies, as there is a corresponding decrease in asteroid ossicles.

Other invertebrate groups are as rare within this biofacies as they are within the tilestones. Rare fragmentary rhynchonellid and terebratulid material is seen within hand specimens and polished sections. Likewise, corals are restricted to derived, rip-up bioclasts of the colonial *Isastraea* sp. and rare specimens of the small, solitary coral *Montlavatia*. Serpulids were recognised in the field as occasional ?*Diptrupa* . Also noted were serpulid-encrusted *Girvillella*. (Metcalf and Underwood, in prep.).

Although terrestrial invertebrates have not been recognised within this biofacies, the biomicrites do contain a well-preserved terrestrial floral element, similar to the tilestones. However, only the more robust elements, such as cycad fruit, conifer cones and indeterminate stems, are common within the coarser shelly beds. These are often confined to the basal portions of large storm beds, but never form the aligned, concentrations that are observed upon bedding planes within the arenaceous tilestone-facies (Metcalf and Underwood, in prep.).

3. *Low diversity mollusc-biofacies II.* This is recognised within the oolites and bioclastic limestones of the Taynton Formation and the coated-grain biomicrites of the Eyford Member seen at Brockhill (Metcalf and Underwood, in prep.).

The oolites contain a very low diversity epifauna dominated by oysters and large pectinids. Oyster debris (*Praeexogyra*) makes up more than 70% of all recognisable bioclasts. The pectinid *Plagiostoma subcardiiformis* is a common macrofaunal element within the oolites and is known to form concentrations elsewhere within the Taynton Formation sequence (Richardson, 1929). This large, thick-shelled flat clam is the most abundant unbroken and hence identifiable element within the limestones, but fragments of other pectinids are also seen in the field (Metcalf and Underwood, in prep.). Other thick-shelled bivalves, such as astartid fragments, are also fairly abundant. Thin-shelled bivalves are rare or absent within this high-energy biofacies in the Eyford area, although a more diverse assemblage of bivalves has been recorded elsewhere within the Taynton Formation succession (e.g. Horton *et al.*, 1987).

Proceritherium is the only gastropod seen in any concentration within the Taynton Formation oolites, forming lenticular accumulations, although unidentifiable gastropod biodebris occur within this biofacies. The only cephalopod material seen during field investigation was small fragments of belemnite guard (Metcalf and Underwood, in prep.).

A low diversity fauna of brachiopods occur within the Taynton oolites. The rhynchonellid *Kallirhynchia* sp. is abundant at some horizons (Horton *et al.*, 1987) and terebratulids also are found within this sequence. The base of the Taynton Formation is characterised by the "*Rhynchonella*-Bed" which reportedly contains *Kallirhynchia* sp.,

and the terebratulids *Stiphrothyris* sp. and *Epithyris* sp., in association with a much more diverse assemblage of macroinvertebrates and microfossils including molluscs, corals, echinoderm debris and several foraminiferan species (Richardson, 1929).

Disarticulated echinoderm material similar to the other biofacies has been recovered from the coated-grain and oolite facies. Isocrinid, millicrinid ossicles and regular echinoid spines make up the major component of these high-energy assemblages. Other invertebrates are rare or absent within this biofacies and plant material is much less common in the oolitic and coated-grain limestones. However, a well preserved conifer twig was recovered from a clay lens, low down in the Taynton Formation sequence at Huntsman's Quarry (Metcalf and Underwood, in prep.).

4. *Vertebrate faunas.* Vertebrate fossils were obtained by systematic field sampling, study of museum collections and by acid preparation of the various limestone facies (Metcalf and Underwood, in prep.). The Eyford Member vertebrate fauna is similar to that of the 'Stonesfield Slates', although the diversity is much lower than at Stonesfield (cf. Savage, 1963, table 1). No mammals or therapsids have been found in the Eyford tilestones and the recorded diversity of tetrapods for these are eleven genera compared with seventeen from the 'Stonesfield Slates' (Savage, 1963). Although this discrepancy can partially be blamed on collection bias in the last century, as Stonesfield roofing tiles were worked by hand, and large sums of money were offered to workmen by Oxford academics to find tetrapods. It is likely that there is some palaeoecological bias as well, as the Eyford Member tilestones are presently hand-cut at Brockhill and employees frequently find pieces of large bone.

The Taynton Limestone Formation has a similar, though slightly less diverse, fauna than the Eyford Member. This is partly a taphonomic bias as the Taynton oolites represent a higher energy environment and only the most robust elements have survived. The diversity of macro- and microvertebrate remains and their palaeoecological niches represented in the different limestone facies is listed briefly in Table 11.1.

Selachian remains are an important element of vertebrate faunas, with both hybodonts and neoselachians being well represented (Metcalf and Underwood, in prep.). The teeth of *Hybodus* are particularly common, being present within most facies. These are dominated by the simple teeth of *H. grossiconus*, although rarer multicuspate teeth represent at least one other species. Highly ornamented multicuspate teeth are common within most microvertebrate samples. These appear to represent a species of *Polyacrodus*, but the general degree of abrasion and lack of roots makes identification uncertain. Other hybodont remains are represented by the large crushing teeth of *Asteracanthus magnus* (up to 40 millimetres in length) and of the smaller *A. tenuis* which also seem to be present within most facies, these species are considered to be synonyms. Rarer hybodont teeth are represented by *Acrodus* from the tilestones (only recorded in museum collections) and small *Lissodus leidodus* from the

biomicrites. Hybodont spines are uncommon, but several taxa have been recorded from museum collections. The commonest of these spine species, *Hybodus apicalis*, is probably synonymous with the most common tooth species, *H. grossiconus*. Hybodont dermal denticles are also encountered within most acid residues and are represented by conical and rare compound forms.

Many acid residues contain a range of small neoselachian teeth (Metcalf and Underwood, in prep.). The commonest of these are referable to the primitive ray *Spathobatis*, a genus present throughout the British marine Bathonian (sections 11.5-11.6). The other relatively abundant taxon is probably referable to the shark *Squalogaleus*. The teeth are variable in form, with thin crowned anterior teeth and more robust lateral teeth. Some of the more gracile teeth closely resemble those of the genus *Protospinax*, to the extent that species of *Protospinax* named on teeth are considered to belong to *Squalogaleus* by Cappetta (1987). In the Types of both genera (named on articulated specimens) only *Squalogaleus* shows the degree of heterodonty seen here. The fin spine *Breviacanthus* from Stonesfield is structurally, but not morphologically, almost identical to that of *Protospinax* (Maisey 1976a) and reveals the existence of a 'Protospinacid' within the tilestone facies, and rare asterate dermal denticles from acid residues are almost identical to those on the Type of *Protospinax* (figured in Maisey 1976b). Rare teeth similar to the anterior teeth of *Squalogaleus*, but with a well developed multicuspate occlusal ridge, could either belong to a second species of *Squalogaleus*, or possibly to *Protospinax ss.*, as no robust lateral teeth were recognised here (Metcalf and Underwood, in prep.).

Other less common neoselachian teeth suggest the presence of some distinctly modern taxa. A number of teeth of a scyliorhinid, possibly *Scyliorhinus*, were recovered from the biomicrites and '*Rhynchonella*-Bed'. These are significantly earlier than any other published scyliorhinid teeth, although *Scyliorhinus* is found at Watton Cliff (section 11.6). A single broken and abraded tooth form the biomicrites possibly represents an orectolobid, but is specifically indeterminate.

In contrast to selachians, chimaeroid remains are rare. The jaw bones appear to be all *Ischyodus*, with palatine plates and a mandibular probably representing two species, with the (?cogenetic) fin spine *Leptacanthus* representing a very rare element.

Disassociated teeth and scales of various holosteans constitute the most abundant fossils within both meso- and microfossil fractions. Although a number of the fish jaw bones and gular plates are assigned to specific taxa, the vast majority of material is generically indeterminate (Metcalf and Underwood, in prep.). In many old collections, however, this material has been assigned to several 'bucket genera', especially *Eomesodon* (pycnodont teeth), *Lepidotes* (ornamented bones and heavy scales) and *Pholidophorus* (light bones).

The commonest macroscopic osteichthyan teeth are those of oval or hemispherical crushing type. Although generally undiagnostic, occasional gular plates allow the recognition of several taxa. The commonest of these is the pycnodont *Eomesodon bucklandi*, although *Microdon* and at least one semionotid are also present. Microvertebrate samples contain gular plate and mandible fragments of a far smaller fish, probably a pycnodont, with closely spaced crushing teeth.

Small pointed teeth are extremely common in many acid residues. These are dominated by teeth with a long, conical root and a spearhead-shaped tip of clear enamel, referred to as 'fish-type I': Metcalf and Underwood, in prep.). Less common are curved teeth, round in section with a smaller enamel tip and strong surface ribbing, known as 'fish-type II' (Metcalf and Underwood, in prep.). Fragments of jaws with fused, stud-like teeth suggest the presence of at least two more taxa. It is possible to infer from the variety of fish material present that the original diversity of fish was high, and probably included several primitive teleosts. The greater number of osteichthyans recorded from Stonesfield is due to a greater abundance of complete jaw bones allowing specific identification, and is thus more due to taphonomic rather than original palaeoecological criteria.

Crocodilian remains are relatively abundant within all of the facies studied, with shed teeth representing the most abundant of tetrapod fossils (Metcalf and Underwood, in prep.). All of the crocodilian material appears to be from marine teleosaurs, and is probably restricted to the genera *Teleosaurus* and *Steneosaurus*. All of the diagnostic skeletal remains (several jaws and jaw fragments and a partial skull) are from the Stonesfield species *Teleosaurus subulidens* although, as at Stonesfield (Phillips 1871), occasional slender teeth may represent a more gracile species (*?T. geoffroyi*). *Steneosaurus* ?*brevidens* is represented by a number of robust, conical teeth, although no cranial material is known. A range of undiagnostic crocodilian remains are also known. These include vertebrae, ribs, limb bones and scutes. Skeletal material has not been recorded in the Taynton Limestone, although this is most likely due to collection failure.

A single Chelonian carapace (BMNH R2634) is recorded from Huntsman's Quarry. This is probably referable to the Bathonian genus *Protochelys*. Several teeth in the Witts collection appear to belong to plesiosaurs, but no skeletal remains are known. Other reptilian remains are more cryptic. A straight tooth (BRSUG.9029-3) from the tilestones, oval in section, bears no resemblance to known crocodilian material, and may represent a rare crocodile (possibly a metriorhynchid cf. Massare, 1987, fig. 5, p.124) or plesiosaur. A heavily abraded large, conical tooth from the Taynton Limestone may belong to a pliosaur, as a well preserved pliosaur tooth is present amongst uncatalogued Stonesfield material in Oxford Museum.

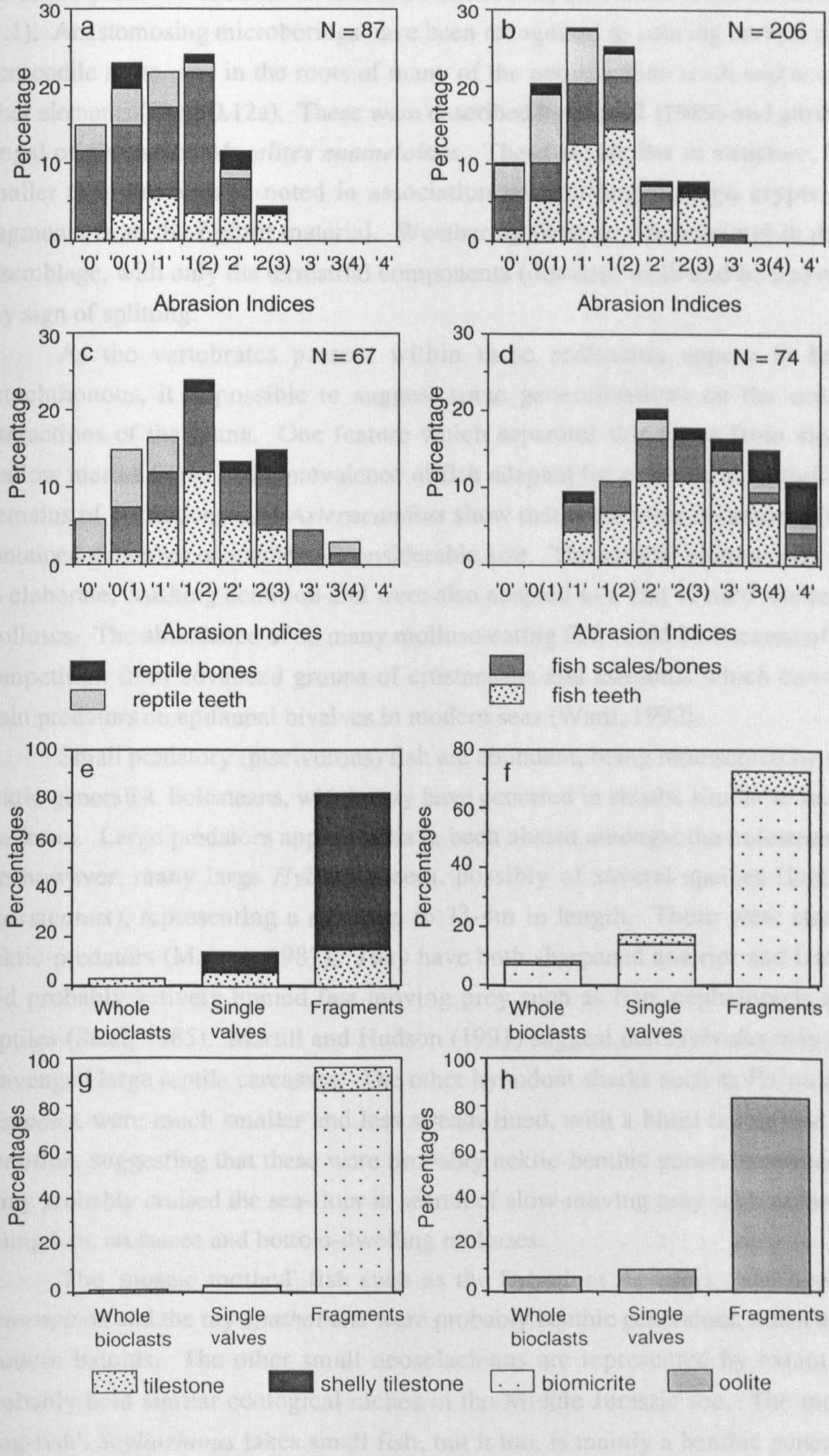
Remains of pterosaurs are not uncommon fossils, with pieces of limb bones and phalanges being amongst the commonest reptile bones (Metcalf and Underwood, in prep.). These remains have traditionally been assigned to *Rhamphocephalus bucklandi* (e.g. Savage, 1963), a species described on fragmentary remains from Stonesfield. Despite this, a well preserved skull cast of *R. prestwichi* has been collected from the Eyford Member (Woodward, 1894) suggesting that several species may be present. A new pterosaur taxon may be represented by very small bones, as yet undescribed, from an undisclosed Eyford Member locality (D. Dartnall, pers. comm.). Although, definite pterosaur teeth have been identified, it is likely that older identifications are unreliable, with confusion between pterosaur and gracile *Teleosaurus* teeth.

Occasional finds of pieces of large reptile bone have been made in many of the Eyford Member quarries. These have by analogy with Stonesfield been generally referred as *Megalosaurus bucklandi*. Although much of this material is indeterminate, some limb bones and teeth may be justifiably referred to *Megalosaurus*. Although no other dinosaur remains have been documented (Spencer & Benton, 1995), an uncatalogued spatulate cetiosaur tooth is present in the Royce collection. This is embedded in a white oolite and presumably came from the Taynton Limestone Formation.

11.7.4. Taphonomy and palaeoecological conclusions

In all I studied over 400 vertebrate remains and a similar amount of invertebrates and plant fragments from the sedimentary facies defined above. The Charlbury and Taynton Formation assemblage differs from all of the other accumulations in not being concentrated within a small area, but rather it is scattered throughout the sections in the quarries studied. None of the vertebrate remains are associated, and the majority exhibit fragmentation and a great deal of sedimentary rounding (Fig. 11.3a-d). Spiral fractures and delicate processes have been lost in most of the meso- and macro- vertebrate fraction through abrasion (Fig. 11.3a-d), but many of the tiniest elements are in much better condition. The degree of rounding exhibited by the vertebrate remains is largely dependent upon the lithofacies in which it occurred (Fig. 11.3a-d). Much more delicate material (fish scales and bones) are preserved in the tilestone facies of the Charlbury Formation (Fig. 11.3a-b) than in the higher-energy lithofacies (Fig. 11.3c,d), and remains tend to be less abraded. Only the most robust elements are preserved in the Taynton oolites (Fig. 11.3d) and all of these show severe spiral fragmentation and are strongly abraded. The tilestones and coarse-grained biomicrite facies (Fig. 11.3e,f) also contain more articulated or whole bioclasts than the higher energy sediments (Fig. 11.3g,h).

Fig. 11.3. (a-d) The degrees of mechanical rounding of vertebrate material in the Eyford lithofacies: (a) Charlbury Formation, "Tilestone" facies; (b) Charlbury Formation, "Tilestone" facies (Witts Collection); (c) Charlbury Formation, Biomicrite facies; (d) Taynton Formation, Biooosparites. Categories in parenthesis indicate an element beginning to display the erosion features of the next abrasion stage. (e-h) The disarticulation and fragmentation of bioclastic material in the Eyford lithofacies: (e) Charlbury Formation, "Tilestone" facies; (f) Charlbury Formation, Biomicrite facies; (g) Charlbury Formation, Coated-grain Biomicrite facies; (h) Taynton Formation, Biooosparites.


a
N = 87
b
N = 206
c
N = 67
d
N = 74
Percentage
30
20
10
0
'0' '0(1)' '1' '1(2)' '2' '2(3)' '3' '3(4)' '4'
Abrasion Indices
reptile bones
reptile teeth
fish scales/bones
fish teeth
e
f
g
h
100
80
60
40
20
0
Percentages
Whole bioclasts
Single valves
Fragments
tilestone
shelly tilestone
biomicrite
oolite

There was no evidence that any of the vertebrate skeletal remains studied had been gnawed, bitten, trampled or digested by another vertebrate. However, endolithic traces are present in both bone and shell substrates, recovered from all facies (Table 11.1). Anastomosing microborings have been recognised as causing surface damage to a crocodile scute and in the roots of many of the neoselachian teeth and occasionally other elements (Fig. 10.12a). These were described by Martill (1989) and attributed to a fungal origin named *Mycelites enameloides.* These are similar in structure, but much smaller than the borings noted in association with *Lithographaga* crypts in oyster fragments from the Eyford material. Weathering damage was minimal in the Eyford assemblage, with only the terrestrial components (dinosaur teeth and bones) exhibiting any sign of splitting.

As the vertebrates present within these sediments appear to be largely autochthonous, it is possible to suggest some generalisations on the ecology and interactions of the fauna. One feature which separates this fauna from any modern shallow marine fauna is the prevalence of fish adapted for a diet of hard-shelled items. Remains of pycnodonts and *Asteracanthus* show that these were common, diverse and contained a number of species of considerable size. The smaller chimaeroids also have an elaborate, crushing dentition and were also adapted to a diet of hard-shelled benthic molluscs. The abundance of so many mollusc-eating fish could be because of a lack of competition from advanced groups of crustaceans and asteroids which constitute the main predators on epifaunal bivalves in modern seas (Ward, 1992).

Small predatory (piscivorous) fish are abundant, being represented by the active nektic-generalist holosteans, which may have occurred in shoals, similar to modern-day mackerel. Large predators appear to have been absent amongst the holosteans. There are however, many large *Hybodus* teeth, possibly of several species (including *H. grossiconus*), representing a shark up to ?3-4m in length. These were stream-lined nektic-predators (Maisey, 1987). They have both sharpened anterior and lateral teeth, and probably actively hunted fast-moving prey such as fish, cephalopods and small reptiles (Steel, 1985). Martill and Hudson (1991) suggest that *Hybodus* may have also scavenged large reptile carcasses. The other hybodont sharks such as *Polyacrodus* and *Lissodus*, were much smaller and less stream-lined, with a blunt lateral and posterior dentition, suggesting that these were probably nektic-benthic generalists (Steel, 1985). They probably cruised the sea-floor in search of slow-moving prey such as the mollusc-eating fish, crustacea and bottom-dwelling molluscs.

The 'mosaic toothed' fish such as the hybodont *Acrodus* , the neoselachian *Protospinax* and the ray *Spathobatis* were probably benthic generalists, much as in many modern batoids. The other small neoselachians are represented by extant taxa and probably held similar ecological niches in the Middle Jurassic sea. The modern day 'dog-fish', *Scyliorhinus* takes small fish, but it too, is mainly a benthic generalist with

crustaceans, worms and molluscs forming the main part of its diet. The dog-fish is a cosmopolitan selachian, with species found in the Atlantic, Indian and Pacific Oceans, up into quite high latitudes around northern Europe and the United States eastern seaboard (Steel, 1985). The neoselachian assemblage also contains a possible orectolobid, the group which contains the 'nurse-' and 'carpet-' sharks, these today, are relatively inoffensive benthic dwellers, which lie inactive on or half-buried in the sea-floor preying upon unwary fish and crustacea (Steel, 1985).

The crocodilians appear to have been fish-eaters, with the difference in dentition between species suggesting a range of diets (Massare, 1987). The more robust nature of the jaws and teeth of *Steneosaurus* suggests a diet of larger fish and possibly carrion, whilst the fish-spear like pattern of more slender, forward pointing teeth of *Teleosaurus* suggest a diet of smaller or faster prey. This same fish-spear pattern of jaws was also possessed by *Rhamphocephalus*, and has been considered suitable for fishing on the wing (Wellnhofer, 1991).

The presence of the therapod dinosaur *Megalosaurus bucklandi* is somewhat enigmatic. This is generally considered to be an allochthonous terrestrial element (Savage, 1963), but is a persistent element within many Bathonian shallow marine deposits, often being the only dinosaur present either as recorded bone material (Spencer & Benton, 1993) or from fossil footprint evidence (Sarjeant, 1974, Andrews & Hudson, 1984). The same is true for the shallow marine Callovian record in France (E. Buffetaut, pers. comm.). In the more terrestrial deposits of the earliest and upper Bathonian, *Megalosaurus* remains are less common than those of sauropods and stegosaurs (this thesis). This possibly suggests that at least one species of *Megalosaurus* (probably *M. bucklandi*) was a locally derived, coastal animal, possibly feeding on the many amphibious but egg-laying crocodiles, turtles and other marine reptiles, and fish stranded upon tidal mud flats.

11.8. Hornsleasow versus other Bathonian microvertebrate sites; general conclusions

In general, the taphonomic studies of the four microvertebrate accumulations show that preservation of skeletal remains tend to reflect sedimentary depositional environment (Table 11.1). For instance, the increased abrasion of derived or reworked fossils in high energy environments, supports the model that transportation causes abrasion of vertebrate material. However, skeletal composition does not necessarily reflect the amount of transportation; although robust elements tend to have a better preservation potential than delicate remains. As both weathered and acid digested material was observed in the transported terrestrial assemblages (Kirtlington and Watton Cliff), it would seem that neither of these processes inhibit the potential of remains to be transported or cause elements to be destroyed by fragmentation or abrasion at a faster rate than non-eroded material.

The skeletal remains from the quiet lagoonal Kirtlington assemblage are much less fragmented or bioeroded than those from Hornsleasow, indicating that biological activity is also an extremely important taphonomic process. As expected the fully marine Eyford assemblage do not show signs of biological activity, other than the profuse amount of bioerosion by micro-organisms.

As with the Hornsleasow assemblage the same problems were encountered in assessing taphonomic and palaeoecological factors for the vertebrate remains, particularly in ascertaining such parameters as fracture differentiation in the microvertebrate fractions. This was particularly true for Watton Cliff and Kirtlington Mammal Bed were only microvertebrate remains are preserved. The lack of contemporary large vertebrate remains within these deposits means that the weathering and abrasion profiles are somewhat untested. However, this size sorting in these accumulations must reflect certain taphonomic or palaeoecological conditions prevailed.

11.9. Specific Research Aim 3: Future microvertebrate taphonomic research

There are two ways in which this research could be developed further. The first would be to carry out many more taphonomic investigations on similar aged localities. In the Bathonian of Great Britain there are a large number of vertebrate-bearing sites (Chapter 1) and the majority of these have not been studied in taphonomic detail. Foreign sites which have yielded terrestrial vertebrate remains include the Guimarota lignite mine (Callovian-Oxfordian) in Portugal. The taphonomy of more open marine Middle Jurassic facies (such as the Oxford Clay faunas) could also be useful for comparative purposes, and an extensive database of the taphonomy and palaeoecology of Middle Jurassic vertebrate-yielding sites could be constructed. Such a database would significantly add to our knowledge of the ancient environment of this part of the fossil record, as well as bringing a greater understanding of the taphonomic effects on similar vertebrate material across the different facies. Working only with Middle Jurassic faunas would lessen the confusion which might be introduced when applying preservational information deduced from modern vertebrate material or Tertiary palaeocommunities to the extinct assemblages of the ancient deposits.

A further method of checking the conclusions derived from this study, would be to study vertebrate and other fossil preservation in similar palaeoenvironments from different ages. Such crocodile infested ponds are known from the Tullock Formation (Palaeocene) of Wyoming (Fiorillo and Hutchison, 1992) and several paleosol horizons have yielded scattered vertebrate remains (Chapter 7). A taphonomic analysis could also be made of a modern crocodile pond or similar karstic pool after careful sampling or possible coring of the sediments had been made. Such a novel approach might answer some of the problems encountered in my study concerning how and when the fossil remains were broken, the rates of burial of such material, any preferential preservational

factors (such as size sorting or micro-environmental parameters) and the applicability of the standard taphonomic theories to such an accumulation.

A. B. 1757. Further account of fossils. *Gentleman's Magazine*, 27: 122-123.

Adams, A.E., MacKenzie, W.S. & Guilford, C. 1984. *Atlas of sedimentary rocks under the microscope*. Longman Scientific & Technical.

Alexandersson, E.T. 1972. Intergranular growth of marine aragonite and Mg calcite: Evidence of precipitation from supersaturated seawater. *Journal of Sedimentary Petrology*, 42: 441-460.

Allen, J.R.L. 1984. Experiments on the settling, overturning and entrainment of bivalve shells and related models. *Sedimentology*, 31: 227-250.

--------- 1986. Pedogenic calcretes in the Old Red Sandstone facies (Late Silurian-early Carboniferous) of the Anglo-Welsh area, southern Britain. *In* (Wright, P.V.; ed.) *Paleosols: their recognition and interpretation*. Blackwells, Oxford, p.58-96.

--------- 1991. Transport - hydrodynamics. Shells. *In* (Briggs, D.E.G. & Crowther, P.R.; eds.) *Palaeobiology: A synthesis.* Blackwell Scientific, Oxford, 227-230.

Allen, P.A. 1975. Wealden of the Weald: a new model. *Proceedings of the Geologists' Association*, 86: 389-426.

Allison, P.A. 1990. Pyrite. *In* (Briggs, D.E.G. & Crowther, P.R.; eds.) *Palaeobiology: A synthesis.* Blackwell Scientific, Oxford, 253-255.

Allison, P.A. & Briggs, D.E.G. 1991. *Taphonomy: releasing the data locked in the fossil record. Topics in Geobiology, Vol. 9.* Plenum, New York.

Anderton, R., Bridges, P.H., Leeder, M.R. & Sellwood, B.W. 1979. *A Dynamic Stratigraphy of the British Isles : A Study in Crustal Evolution.* Allen & Unwin, London, 301pp.

Andrews, J.E. 1985. Sedimentary facies of a late Bathonian regressive episode: the Kilmaluag and Skudiburgh formations of the Great Estuarine Group, Inner Hebrides, Scotland. *Journal of the Geological Society of London*, 142: 1119-1137.

--------- & Hudson, J.D. 1984. First Jurassic dinosaur footprint from Scotland. *Scottish Journal of Geology*, 20: 129-134.

--------- & Walton, W. 1990. Depositional environments within Middle Jurassic oyster-dominated lagoons: an integrated litho-, bio- and palynofacies study of the Duntulm Formation (Great Estuarine Group, Inner Hebrides). *Transactions of t he Royal Society of Edinburgh: Earth Sciences*, 81: 1-22.

Andrews, P. 1990. *Owls, caves and fossils*. University of Chicago Press, Chicago.

--------- & Cook, J. 1985. Natural bone modifications to bones in a temperate setting. *Man*, 20: 675-691.

--------- & Nesbit-Evans, E. 1983. Small mammal bone accumulations produced by mammalian carnivores. *Paleobiology*, 9: 289-307.

Antia, D.D.J. 1979. Bone-beds: A review of their classification, occurrence, genesis, diagenesis, geochemistry, palaeoecology, weathering, and microbiotas. *Mercian Geologist*, 7: 93-174.

Argast, S., Farlow, J.O., Gabet, R.M. & Brinkman, D.L. 1987. Transport-induced abrasion of fossil reptilian teeth: implications for the existence of Tertiary dinosaurs in the Hell Creek Formation, Montana. *Geology*, 15: 927-930.

Arkell, W.J. 1931. The Upper Great Oolite, Bradford Beds and Forest Marble of south Oxfordshire and the succession of gastropod faunas in the Great Oolite. *Quarterly Journal of the Geological Society of London*, 87:563-629.

-------- , 1933. *The Jurassic System in Great Britain*. Oxford University Press, Oxford. 681pp.

--------- 1947a. *The Geology of Oxford.* Oxford. 267pp.

--------- 1947b. *Oxford stone*. Faber & Faber, London. 185pp.

--------- 1951. The geology of the Corallian ridge between Wootton Bassett and Lyneham, Wiltshire. *Wiltshire Archaeological natural History Magazine*, 54: 1-18.

--------- 1951-1958. A monograph of English Bathonian ammonites. *Monograph of the Palaeontographical Society of London*, 8: 264pp.

--------- 1956. *Jurassic Geology of the World.* Oliver & Boyd, Edinburgh. 806pp.

--------- & Donovan, D.T. 1952. The Fuller's Earth of the Cotswolds and its relation to the Great Oolite. *Quarterly Journal of Geological Society of London*, 107: 227-253.

Arndorff, L. 1993. Lateral relations of deltaic palcosols from the Lower Jurassic Rønne Formation on the island of Bornholm, Denmark. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 100: 235-250.

Ashford, S. 1992. *Preservation and derivation in a mid Jurassic paleosol.* Unpublished BSc thesis, University of Bristol, 45pp.

Aston, M.A. 1974. Stonesfield Slate. *Department of Museum Services of Oxfordshire County Council*, Publication 5: 86pp.

Audley-Charles, M.G. 1970a. Stratigraphical correlation of the Triassic rocks of the British Isles. *Quarterly Journal of Geological Society of London*, 126: 19-48.

--------- 1970b. Triassic palaeogeography of the British Isles. *Quarterly Journal of Geological Society of London*, 126: 49-849.

Baas-Becking, L.G.M., Kaplan, I.R. & Moore, D. 1960. Limits of the natural environment in terms of pH and oxidation-reduction potentials. *Journal of Geology*, 68: 243-284.

Badgley, C. 1986. Counting individuals in mammalian fossil assemblages from fluvial environments. *Palaios* 1: 328-338.

Baker, P.G. 1981. Interpretation of the Oolite Marl (Upper Aalenian, Lower Inferior Oolite) of the Cotswolds, England. *Proceedings of the Geologists Association*, 92:169-187.

Barford, N.C. 1985. *Experimental measurements : precision, error and truth*. John Wiley & Sons, Chichester.

Barnard, P.C. & Cooper, B.S. 1981. Oil and source rocks of the North Sea area. In (Illing, L.V. & Hobson, G.D.; eds) *Petroleum Geology of the Continental Shelf of North-West Europe*. Heyden, London, 432-443.

Barnard, P.D.W. 1973. Mesozoic floras. *In* (Hughes, N.F.; ed.) *Organisms and Continents through Time*. Special Papers in Palaeontology, 12: 175-188.

Bartisiokas, A. & Middleton, A.P. 1992. Characterisation and dating of recent and fossil bone by x-ray diffraction. *Journal of Archaeological Science*, 19: 63-72.

Bate, R.H. 1965. Middle Jurassic Ostracoda from the Grey Limestone Series, Yorkshire. *Bulletin of the British Museum of Natural History*, Series Geological, 11: 73-134.

--------- 1967a. Stratigraphy and palaeogeography of the Yorkshire Oolites and their relationship with the Lincolnshire Limestone. *Bulletin of the British Museum of Natural History*, Series Geological, 14: 111-141.

--------- 1967b. The Bathonian Upper Estuarine Series of eastern England, Part 1 Ostracoda. *Bulletin of the British Museum of Natural History*, Series Geological, 14: 21-66.

--------- , 1978. The Jurassic. Part II Aalenian to Bathonian. In (Bate, R.H. and Robinson, J.E.; eds.) *A stratigraphical index of British Ostracoda. Geological Journal Special Issue*, 8:213-258. Seel House Press, Liverpool.

Bathurst, R.G.C. 1966. Boring algae, micritic envelopes and lithification of molluscan biosparites. *Geological Journal*, 5: 15-32.

--------- 1975. *Carbonate sediments and their diagenesis*. Elsevier, Amsterdam.

Beach, D.K. & Ginsburg, R.N. 1980. Facies succession, Plio-Pleistocene carbonate, north-western Great Bahama Bank. *American Association of Petroleum Geologists Bulletin*, 64: 1634-1642.

Beesley, T. 1877. On the geology of the eastern portion of the Banbury and Cheltenham direct railway. *Proceedings of the Geologists' Association*, 5: 185.

Behrensmeyer, A.K. 1975. The taphonomy and palaeoecology of Plio-Pleistocene vertebrate assemblages of Lake Rudolf, Kenya. *Bulletin of the Museum of Comparative Zoology*, 146: 473-578.

--------- 1978. Taphonomic and ecological information from bone weathering. *Paleobiology*, 4: 150-162.

--------- 1982. Time resolution in fluvial vertebrate assemblages. *Paleobiology*, 8: 211-228.

--------- 1988. Vertebrate preservation in fluvial channels. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 63: 183-199.

--------- 1990. Transport-hydrodynamics: bones. In (Briggs, D.E.G. & Crowther, P.R.; eds.) *Paleobiology: a synthesis*. Blackwell Scientific Publications, Oxford, 232-235.

--------- 1991. Terrestrial vertebrate accumulations. *In* (Allison, P.A. & Briggs, D.E.G.; eds.) *Taphonomy: releasing the data locked in the fossil record.* Topics in Geobiology, vol. 9. Plenum Press, New York, 291-335.

--------- & Dechant-Boaz, D.E. 1980. The recent bones of Amboseli Park, Kenya, in relation to East African paleobiology. *In* (Behrensmeyer, A.K. & Hill, A.P.; eds.) *Fossils in the making: vertebrate taphonomy and palaeoecology*. University of Chicago Press, Chicago, 72-92.

--------- , Gordon, K.D. & Yanagi, G.T. 1986. Trampling as a cause of bone surface damage and pseudo-cutmarks. *Nature*, 319: 768-771.

--------- , --------- , & --------- 1989. Nonhuman bone modification in Miocene fossils from Pakistan. *In* (Bonnichsen, R. & Sorg, M.H.; eds.) *Bone Modification.* Center for the study of the first Americans, Orono, Maine, 99-120.

--------- & Hill, A.P. 1980. *Fossils in the making: vertebrate taphonomy and palaeoecology*. University of Chicago Press, Chicago.

--------- & Hook, R.W. 1992. Paleoenvironmental contexts and taphonomic modes. *In* (Behrensmeyer, A.K., Damuth, J.D., DiMichele, W.A., Potts, R., Sues, H.-D. & Wing, S.L.; eds.) *Terrestrial ecosystems through time*. University of Chicago Press, Chicago, 15-136.

--------- & Kidwell, S.M. 1985. Taphonomy's contribution to paleobiology. *Palaeobiology*, 11: 105-119.

--------- & Schindel, D. 1983. Resolving time in paleobiology. *Paleobiology*, 9: 1-8.

--------- , Western, D. & Dechant-Boaz, D.E. 1979. New perspectives in vertebrate palaeoecology from a recent bone assemblage. *Paleobiology*, 5: 12-21.

Benton M.J. 1988. British fossil reptile sites. *Special Papers in Palaeontology*, 40: 73-84.

--------- 1993a. Reptilia. In (Benton, M.J.; ed.) *The Fossil Record 2*. Chapman & Hall, London, p. 681-716.

--------- 1993b. *The Fossil Record 2*. Chapman & Hall, London, 845pp.

--------- & Spencer, P.S. 1994 (in press). *British Fossil Reptile Sites*. Chapman & Hall, London.

--------- , Metcalf, S.J., Evans, S.E., Milner, A.R., Savage, R.J.G. & Unwin, D.M. 1995 (in preparation). The Middle Jurassic vertebrate faunal assemblage of Hornsleasow, Gloucestershire.

Berner, R.A. 1970. Sedimentary pyrite formation. *American Journal of Science*, 268: 1-23.

--------- 1984. Sedimentary pyrite, an update. *Geochimica et Cosmochimica Acta*, 48: 605-615.

Bernouilli, D & Jenkyns, H.C. 1974. Alpine, Mediterranean and North Atlantic Mesozoic facies in relation to the early evolution of the Tethys. In (Dott, R.H. & Shaver, R; eds.) *Geosynclinal sedimentation, Modern and Ancient, a Symposium*. Society of Economic Paleontologists and Mineralogists, Special Publication, 28:129-160.

Boaz, D. 1980. Fossils in the making: taphonomy and vertebrate palaeoecology. *Current Anthropology*, 21: 404-406.

Boneham, B.F.W. & Wyatt, R.G. 1993. The stratigraphical position of the Middle Jurassic (Bathonian) Stonesfield Slate of Stonesfield, Oxfordshire, U.K. *Proceedings of the Geologists' Association*, 104: 123-137.

Bott, M.H.P. 1974. The geological interpretation of a gravity survey of the English Lake District and the Vale of Eden. *Journal of the Geological Society, London*, 130: 309-328.

Boulter, M.C. & Windle, T. 1993. A reconstruction of some Middle Jurassic vegetation in Northern Europe. *Special Papers in Palaeontology*, 49: 125-154.

Bown, T.M. & Kraus, M.J. 1981. Vertebrate fossil-bearing paleosol units (Willwood Formation, Lower Eocene, north-west Wyoming, U.S.A.) : implications for taphonomy, biostratigraphy and assemblage analysis. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 34: 31-56.

Bradshaw, M.J. 1978. *A facies analysis of the Bathonian of eastern England*. DPhil thesis, University of Oxford.

--------- Cope, J.C.W., Cripps, D.W., Donovan, D.T., Howarth, M.K., Rawson, P.F., West, I.M. & Wimbledon, W.A. 1992. Jurassic. In (Cope J.C.W., Ingham, J. & Rawson, P.S.: eds.) *Atlas of Palaeogeography and Lithofacies*. Geological Society, London, Memoir, 13: 107-129.

Brain, C.K. Bone weathering and the problem of bone pseudo-tools. *South African Journal of Science*, 63: 97-99.

--------- 1981. *The hunters and the hunted? An introduction to African cave taphonomy*. University of Chicago Press, Chicago.

Brand, E. & Hoffman, K. 1963. Stratigraphy and facies of the north-west German Jurassic and genesis of its oil deposits. *Proceedings of the 6th World Petroleum Congress Section 1*, 223-245.

Brandt, D.S. 1989. Taphonomic grades as a classification for fossiliferous assemblages and implications for palaeoecology. *Palaios*, 4: 303-309.

Brett, C.E. 1990. Destructive taphonomic processes and skeletal durability. *In* (Briggs, D.E.G. & Crowther, P.R.; eds.) *Palaeobiology: A synthesis.* Blackwell Scientific, Oxford, 223-226.

--------- & Baird, G.C. 1986. Comparative taphonomy: a key to paleoenvironmental interpretation based on fossil preservation. *Palaios*, 1: 207-227.

Brewer, R. 1976. *Fabric and mineral analysis of soils.* Kreiger Press, New York.

Briggs, D.E.G. 1990. Flattening. *In* (Briggs, D.E.G. & Crowther, P.R.; eds.) *Palaeobiology: A synthesis.* Blackwell Scientific, Oxford, 244-247.

--------- & Crowther, P.R. *Palaeobiology: A synthesis.* Blackwell Scientific, Oxford

Broderip, W.J. 1828. Observations on the jaw of a fossil mammiferous animal found in the Stonesfield Slate. *Zoological Journal of London*, 3: 408-411.

Bromage, T.G. 1984. Interpretation of scanning electron microscopic images of abraded bone surfaces. *American Journal of Physical Anthropology*, 64: 161-178.

Brongniart, A. 1829. *Tableau des terrains qui composent l'ecorce du globe, ou essai sur la structure de la partie connue de la terre.* Paris and Strasbourg, 435pp.

Brookes, R. 1763. *The natural history of waters, earths, stones, fossils, and minerals; with their virtues, properties and medicinal uses; to which is added the method in which Linnaeas has treated these subjects. Volume 5.* Newberry, London.

Brown, D.S. & Keen, J.A. 1991. An extensive marine vertebrate fauna from the Kellaways Sand (Callovian, Middle Jurassic) of Lincolnshire. *Mercian Geologist*, 12: 87-96.

Brunstom, R.G. & Walmsley, P.J. 1969. Permian evaporites in the North Sea Basin. *Bulletin of the American Association of Petroleum Geologists*, 53: 870-883.

Buckland, W. 1818. Table appended to W. Phillips : *Selection of Facts arranged so as to form an Outline of the Geology of England and Wales.* London.

--------- 1824 Notice on the *Megalosaurus*, or great fossil lizard of Stonesfield. *Transactions of the Geological Society of London, Series 2*, 1: 390-396.

--------- 1828 Note sur les traces de tortues observées dans le gres rouge. *Annales de Sciences Naturelles, Paris*, 13: 85-86.

--------- 1836. Geology and mineralogy considered with reference to natural theology. *The Bridgewater Treatises*, 6 (2 volumes), 600pp + 128pp.

Buckman, S.S. 1901. The Bajocian and contiguous deposits in the North Cotteswolds: the main hill-mass. *Quarterly Journal of the Geological Society of London*, 57: 126-155.

Bullock, N., Fedoroff, N., Jongerius, A., Tursina, T. & Babel, U. 1985. *Handbook of soil thin section description*. Waine Research, Albrighton.

Cadée, G.C. 1990. The history of taphonomy. *In* (Donovan, S.K.; ed.) *The processes of fossilisation*. Columbia University Press, New York, 3-21.

Callomon, J.H. 1955. The ammonite succession in the Oxford Clay and Kellaways Beds at Kidlington, Oxfordshire, and the zones of the Callovian Stage. *Philosophical Transactions of the Royal Society of London*. B239:215-264.

--------- 1964. Notes on the Callovian and Oxfordian Stages. *In* (Mauberge, P.L.; ed.) *Colloque du Jurassique a Luxembourg 1962*. Luxembourg, 269-291.

--------- 1979. Marine boreal Bathonian fossils from the northern North Sea and their palaeogeographical significance. *Proceedings of the Geologists' Association*, 90: 163-169.

--------- , Oates, M.J., Coe, A.L. & Metcalf, S.J. 1993. *Day excursion B : Jurassic of Oxfordshire and Cotswolds. Arkell International Symposium on Jurassic Geology.* Geological Society Guide. 47pp.

Canfield, D.E. & Raiswell, R. 1991a. Carbonate precipitation and dissolution: its relevance to fossil preservation. *In* (Allison, P.A. & Briggs, D.E.G.; eds.) *Taphonomy: releasing the data locked in the fossil record.* Plenum Press, New York, 412-445.

--------- & ---------. 1991b. Pyrite formation and fossil preservation. *In* (Allison, P.A. & Briggs, D.E.G.; eds.) *Taphonomy: releasing the data locked in the fossil record.* Plenum Press, New York, 338-382.

Cappetta, H. 1987. Chondrichthyes II; Mesozoic and Cenozoic Elasmobranchii. *Handbook of Paleoichthyology*, 3B: 1-193.

Carroll, R. L. 1988. *Vertebrate paleontology and evolution*. W. H. Freeman, San Francisco.

Cave, R. 1977. *Geology of the Malmesbury District.* Memoirs of the Geological Survey of Great Britain. 343pp.

Channon, P.J. 1950. New and enlarged Jurassic sections in the Cotswolds. *Proceedings of the Geologists' Association*, 61: 242-260.

Charlesworth, E. 1855. Notice on new vertebrate fossils. *Report of the British Association for the Advancement of Science*, 1854 (1855): 80.

Chesworth, W. 1973. The parent rock effect in the genesis of soils. *Geoderma*, 10: 215-225.

Chidlaw, N. 1987. *The sedimentology of the Marlstone Rock Bed and Dyrham Silt Formations (Pliensbachian, Lower Jurassic) of the Cotswolds Hills*. PhD thesis, University of Bristol.

--------- & Campbell, M.J.A. 1988. Sedimentation patterns in the Cornbrash Limestone Formation of the Cotswold Hills, *Proceedings of the Geologists' Association*, 99: 27-42.

Chowdhury, A.N. 1982. Variations in diagenetic environments of the Coral Rag and underlying oolite of the Osmington Oolite Formation (Upper Oxfordian) of the Oxford-Berkshire area, U.K., and their implications. *Sedimentary Geology*, 32:247-263.

Clark, J. & Etches, S. 1991. Predation amongst Jurassic marine reptiles. *Proceedings of the Dorset Natural History and Archaeological Society*, 113: 202-205.

Cohen, A.D. 1974. Evidence of fires in the ancient Everglades and coastal swamps of Southern Florida. *Environments of South Florida, Present and Past. Miami Geological Society Memoirs*, 213-218.

Colbert, E.H. 1964. Climate zonation and terrestrial faunas. *In* (Nairn, A.E.M; ed.) *Problems in Palaeoclimatology*. Interscience, New York, 617-638.

Cole, J. 1989. *Palynofacies report of a sample from a vertebrate bearing clay lens from Hornsleasow Quarry*. Unpublished report, Gerhart Geodata, Aberdeen. 10pp.

Conybeare, W.D. & Phillips, W. 1822. *Outlines of the geology of England and Wales. Part One*. W. Phillips, London, 470pp.

Contini, D., Elmi, S. & Mouterde, R. 1971. Les zones du Jurassique en Iranie : Aalenian. *C. r. Somm. Soc. geol. Fr.* 1971:84-85.

Coogan, A. H. 1971. Bahamian and Floridian Biofacies. Florida Keys and Western Bahamas. *In* (Multer, H.G.; ed.) *Field Guide to Carbonate Rock Environments*. 141-152.

Cope, J.C.W., Duff, K.L., Parsons, C.F., Torrens, H.S. Wimbledon, W.A. & Wright, J.K., 1980a. *A correlation of Jurassic rocks in the British Isles, Part Two : Middle and Upper Jurassic*. Geol. Soc. Lond. Special Report, 15. Blackwell Scientific Publications, Oxford. 107pp.

--------- , Getty, T.A., Howarth, M.K., Morton, N. & Torrens, H.S. 1980b. *A correlation of Jurassic rocks in the British Isles, Part One : Introduction and Lower Jurassic*. Geol. Soc. Lond. Special Report, 15. Blackwell Scientific Publications, Oxford. 73pp.

---------- , Ingham, J. & Rawson, P.S. 1992 (eds.) *Atlas of Palaeogeography and Lithofacies*. Geological Society, London, Memoir, 13.

Cope, M.J. 1980. Physical and chemical properties of coalified and charcoalified phytoclasts from some British Mesozoic sediments: an organic geochemical

approach to paleobotany. *In* (Douglas, A.G. & Maxwell, J.R.; eds.) *Advances in organic geochemistry 1979*. Pergamon Press, Oxford, 663-677.

--------- 1993. A preliminary study of charcoalified plant fossils from the Middle Jurassic Scalby Formation of North Yorkshire. *Special Papers in Palaeontology*, 49: 101-111.

Cox, A.H. & Trueman, A.E. 1920. Intra-Jurassic movements and the underground structure of the southern Midlands. *Geological Magazine*, 57:198-208.

Curray, J.R. 1956. The analysis of two-dimensional orientation data. *Journal of Geology*, 64: 117-131.

Currie P. J. 1987. Bird-like characteristics of the jaws and teeth of troodontid theropods (Dinosauria, Saurischia). *Journal of Vertebrate Paleontology*, 7: 72-81.

--------- , Rigby, J. K., Jr., and Sloan, R. E. 1990. Theropod teeth from the Judith River Formation of southern Alberta, Canada. *In* (Carpenter, K. & Currie, P. J.; eds.) *Dinosaur systematics*. Cambridge University Press, Cambridge, 107-126.

Curtis, C.D. 1985. Clay mineral precipitation and transportation during burial diagenesis. *Philosophical Transactions of the Royal Society of London*, A315: 91-105.

Cypert, E. 1972. Plant succession on burned areas in Okefenokee swamp following the fires of 1954 and 1955. *Proceedings of the Annual Tall Timbers Fire Ecology Conference*, 12: 199-217.

Darlington, J. 1988. *Hornsleasow Quarry Archive Report.* Crickley Hill Archaeological Trust, Gloucestershire. 20pp.

Davies, J.R. 1991. Karstification and pedogenesis on a late Dinantian carbonate platform, Anglesey, North Wales. *Proceedings of the Yorkshire Geological Society*, 48: 297-321.

Delair, J.B. & Sarjeant, W.A.S. 1975. The earliest discoveries of dinosaurs. *Isis*, 66: 5-25.

Delair, J.B. & Sargeant, W.A.S. 1985. History and bibliography of the study of fossil vertebrate footprints in the British Isles : Supplement 1973-1983. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 49: 123-160.

Dettman, M.E. 1963. Upper Mesozoic microfloras from S.E. Australia. *Proceedings of the Royal Society of Victoria*, 77: 1-148.

Dewey, J.F., Pitman, W.C., Ryan, W.B.F. & Bonnin, J. 1973. Plate tectonics and the evolution of the Alpine system. *Bulletin of the Geological Society of American*, 84: 3134-3180.

Dickson, J.A.D. 1966. Carbonate identification and genesis as revealed by staining. *Journal of Sedimentary Petrology*, 36: 491-505.

Dietz, R.S. & Holden, J.C. 1970. Reconstruction of Pangaea: Break-up and dispersion of continents, Permian to present. *Journal of Geophysical Research*, 75: 4939-4956.

Dixon, J.E., Fitton, J.G. & Frost, R.T.C. 1981. The tectonic significance of post-Carboniferous igneous activity in the North Sea Basin. In (Illing, L.V. & Hobson, G.D.; eds) *Petroleum Geology of the Continental Shelf of North-West Europe*. Heyden, London, 121-137.

Dodson, P. 1971. Sedimentology and taphonomy of the Oldman Formation (Campanian), Dinosaur Provincial Park, Alberta (Canada). *Palaeogeography, Palaeoclimatology, Palaeoecology*, 10: 21-74.

--------- 1973. The significance of small bones in palaeoecological interpretation. *University of Wyoming Contributions to Geology*, 12: 15-19.

--------- , Behrensmeyer, A.K., Bakker, R.T. & McIntosh J.S. 1980 Taphonomy and paleoecology of the dinosaur beds of the Jurassic Morrison Formation. *Paleobiology*, 6: 208-232.

--------- & Wexlar, D. 1979. Taphonomic investigations of owl pellets. *Paleobiology*, 5: 279-284.

Donn, W.L. 1982. The enigma of high latitude paleoclimate. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 40: 199-212.

Donovan, S.K. 1990. *The processes of fossilisation*. Columbia University Press, New York

Douglas, J.A. and Arkell, W.J. 1932. The stratigraphical distribution of the Cornbrash, II. The north-eastern area . *Quarterly Journal of the Geological Society of London*, 88:112-132.

Duff, K.L. 1975. Palaeoecology of a bituminous shale: the Lower Oxford Clay of central England. *Palaeontology*, 18: 443-482.

Duffin, C. J. 1993. Mesozoic chondrichthyan faunas. 1. Middle Norian (Upper Triassic) of Luxembourg. *Palaeontographica, Abteilung A*, 229: 15-36.

--------- & Delsate, D. 1993. The age of the Upper Triassic vertebrate fauna from Attert (Province of Luxembourg, Belgium). *Belgian Geological Survey, Professional Paper*, 264: 33-44.

Dunham, R.J. 1962. Classification of carbonate rocks according to depositional texture. *In*: (Hann, W.E.; ed.) *Classification of Carbonate Rocks*. Memoirs of the American Association of Petroleum Geologists, Tulsa, 108-121.

Efremov, I.A. 1940. Taphonomy: a new branch of palaeontology. *Pan-American Geologist* 74: 81-93.

Elliot, R.E. 1985. Quantification of peat to coal compaction stages, based especially on phenomena in the East Pennine coal field, England. *Proceedings of the Yorkshire Geological Society*, 45: 163-172.

Ensom, P.C., 1977. A therapsid tooth from the Forest Marble (Middle Jurassic) of Dorset. *Proceedings of the Geologists' Association,* 88: 201-205.

--------- 1994. A lower molar of *Stereognathus* sp. (Reptilia, Therapsida) from the Bathonian of southern England. *Proceedings of the Dorset Natural History and Archaeological Society*, 115:139-141.

--------- , Evans, S.E., Francis, J.E., Kielan-Jaworowska, Z, & Milner, A.R. 1994. The fauna and flora of the Sunnydown Farm footprint site and associated sites: Purbeck Limestone Formation, Dorset. *Proceedings of the Dorset Natural History and Archaeological Society,* 115:181-182.

Esteban, M. & Klappa, C.F. 1983. Subaerial exposure environment. In (Scholle, P.A., Bebout, D.G. & Moore, C.H.; eds.) *Carbonate depositional environments. Memoirs of the American Association of Petroleum Geologists*, 33:1-54.

Evans, S.E., 1989. New material of *Cteniogenys* (Reptilia:Diapsida; Jurassic) and a reassessment of the phylogenetic position of the genus. *Neues Jarbuch fur Geologie und Paläontologie, Monatshefte*, 10: 577-587.

--------- , 1990. The skull of *Cteniogenys*, a choristodere (Reptilia: Archosauromorpha) from the Middle Jurassic of Oxfordshire. *Zoological Journal of the Linnean Society,* 99: 205-237.

--------- , 1991. A new lizard-like reptile (Diapsida: Lepidosauromorpha) from the Middle Jurassic of England. *Zoological Journal of the Linnean Society,* 103: 391-412.

--------- , 1992. Small reptiles and amphibians from the Forest Marble (Middle Jurassic) of Dorset. *Proceedings of the Dorset Natural History and Archaeological Society*, 113:201-202.

--------- & Milner, A.R. 1994. Middle Jurassic microvertebrate faunas from the British Isles. *In* (Fraser, N.C. & Sues, H.-D.; eds.) *In the shadow of the dinosaurs: early Mesozoic tetrapods.* Cambridge University Press, New York, 303-321.

--------- , --------- & Musselt, F. 1988. The earliest known salamanders (Amphibia, Caudata) : a record from the Middle Jurassic of England. *Geobios* , 21: 539-552.

--------- , --------- , & ---------. 1990. A discoglossid frog from the Middle Jurassic of England. *Palaeontology,* 33: 299-311.

Eynon, G. 1981. Basin development and sedimentation in the Middle Jurassic of the northern North Sea. *In* (Illing, L.V. & Hobson, G.D.; eds.) *Petroleum Geology of the Continental Shelf of North-West Europe*. Heyden, London, 196-204.

Fagerstrom, J.A. 1964. Fossil communities in palaecoecology: their recognition and significance. *Geological Society of America Bulletin*, 75: 1197-1216.

Farlow, J. O., Brinkman, D. L., Abler, W. L., & Currie, P. J. 1991. Size, shape, and serration density of theropod dinosaur lateral teeth. *Modern Geology,* 16: 161-198.

Fairchild, I. 1983. Chemical control of cathodoluminescence of natural dolomites and calcite: new data and review. *Sedimentology*, 30:579-583.

--------- 1988. Chemical analysis of sedimentary rocks. *In* (Tucker, M.E.; ed.) *Techniques in Sedimentology*. Blackwell Scientific Publications, Oxford. 234-267.

Fenton, J.P.G., Riding, J.B. & Wyatt, R.J. 1994. Palynostratigraphy of the Middle Jurassic 'White Sands' of central England. *Proceedings of the Geologists' Association*, 105: 225-231.

Finks, R.M. 1979. Fossils and fossilisation. *In* (Fairbridge, R.W. & Jablonski, D.; eds.) *Encyclopedia of paleontology*. Dowden, Hutchison & Ross Inc., Stroudsburg, 327-332.

Fiorillo, A.R. 1984. An introduction to the identification of trample marks. *Current Research*, 1: 47-48.

--------- 1987. Trample marks: caution from the Cretaceous. *Current Research in the Pleistocene*, 4: 73-75.

--------- 1988a. A proposal for a graphic representation of orientation data from fossils. *Contributions in Geology, University of Wyoming*, 26: 1-4.

--------- 1988b. Aspects of bone modification applied to time resolution in the fossil record - an example from the Miocene of Western Nebraska. *Current Research in the Pleistocene, Special Focus : Bone Modification*, 5: 103-109.

--------- 1988c. Taphonomy of Hazard Homestead Quarry (Ongallala Group), Hitchcock County, Nebraska. *Contributions to Geology University of Wyoming*, 26: 57-97.

--------- 1989. An experimental study of trampling: implications for the fossil record. *In* (Bonnichsen, R. & Sorg, M.H.; eds.) *Bone modification*. University of Maine, Orono, 61-71.

--------- 1991. Prey bone utilisation by predatory dinosaurs. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 88: 157-166.

--------- & Hutchison, J.H. 1992. Evidence of pond making by vertebrates in the Tullock Formation. *Journal of Vertebrate Palaeontology*, 12 Abstracts, 28A.

Fisher, D.C. 1981 Crocodilian scatology, microvertebrate concentrations and enamel-less teeth. *Paleobiology*, 7: 262-275.

Fitton, W.H. 1828. The strata from whence the fossil described in the preceding notice was obtained. *Zoological Journal*, 3: 412-418.

FitzPatrick, E.A. 1984. *Micromorphology of soils*. Chapman & Hall, London.

--------- 1993. *Soil Microscopy and micromorphology*. John Wiley & Sons, Chichester.

Folk, R.L. 1959. Practical petrographic classification of limestone types. *Bulletin of American Association of Petroleum Geologists'*, 43: 1-38.

--------- , 1962. Spectral subdivision of limestone types. *In*: (Hann, W.E.; ed.) *Classification of Carbonate Rocks*. Memoirs of the American Association of Petroleum Geologists, Tulsa, 62-84.

--------- & Ward, W. 1957. Brazos River bar: a study in the significance of grain size parameters. *Journal of Sedimentary Petrology*, 27:3-26.

Food and Agricultural Organisation (F.A.O.) 1971-1981. *Soil map of the world, 1 : 5 000 000, 10 volumes*. United Nations Educational, Scientific and Cultural Organisation (U.N.E.S.C.O.), Paris.

Foos, A.M. 1991. Aluminous lateritic soils, Eleuthera, Bahamas: A modern analog to carbonate paleosols. *Journal of Sedimentary Petrology*, 61:340-348.

Ford, D.C. & Williams, P.W. 1989. *Karst geomorphology and hydrology*. Unwin Hyman, London.

Fox-Strangeways, C. 1903. *The Geology of the Country near Leicester* (Sheet 156). memoirs of the Geological Survey, HMSO, London.

Frakes, L.A. 1979. *Climates throughout Geologic Time*. Elsevier, Amsterdam.

Frank, J.R., Carpenter, A.B. & Oglesby, T.W. 1982. Cathodoluminescence and composition of calcite cement in the Taum Sauk limestone (Upper Cambrian), southeast Missouri. *Journal of Sedimentary Petrology*, 52:631-638.

Fraser, N.C. 1985. Vertebrate faunas from Mesozoic fissure deposits of south-west Britain. *Modern Geology*, 9: 273-300.

--------- 1986. Triassic sphenodontids from south-west England and a review of their classification. *Palaeontology*, 29: 165-186.

--------- Rare tetrapod remains from the late Triassic fissure infillings of Cromhall Quarry, Avon. *Palaeontology*, 31: 567-576.

Freakes, C.R. & Retallack, G.J. 1988. Recognition and chemical characterisation of fossil soils developed on alluvium; a Late Ordovician example. *In* (Reinhardt, J. & Sigleo, W.R.; eds.) *Paleosols and weathering through time: principles and applications. Geological Society of America Special Papers*, 216: 35-48.

Freeman, E.F., 1976. Mammal teeth from the Forest Marble (Middle Jurassic) of Oxfordshire, England. *Science*, 194: 1053-1055.

--------- , 1979. A Middle Jurassic mammal bed from Oxfordshire. *Palaeontology*, 22: 135-166.

Frey, M. 1987. *Low temperature metamorphism*. Blackie, Glasgow.

Fürsich, F.T. 1977. Corallian (Upper Jurassic) marine benthic associations from England and Normandy. *Palaeontology*, 20: 337-386.

--------- 1978. The influence of faunal condensation and mixing on the preservation of fossil benthic communities. *Lethaia*, 11: 243-250.

Galton, P. M. 1975. English hypsilophodontid dinosaurs (Reptilia: Ornithischia). *Palaeontology*, 18: 741-752.

--------- 1976. *Iliosuchus*, a Jurassic dinosaur from Oxfordshire and Utah. *Palaeontology*, 19: 587-589.

--------- 1980. Armored dinosaurs (Ornithischia: Ankylosauria) from the Middle and Upper Jurassic of England. *Geobios*, 13: 825-837.

--------- 1985. British plated dinosaurs (Ornithischia, Stegosauridae). *Journal of Vertebrate Paleontology*, 5: 211-254.

--------- & Powell, H. P. 1983. Stegosaurian dinosaurs from the Bathonian (Middle Jurassic) of England, the earliest record of the Family Stegosauridae. *Geobios*, 16: 219-229.

Gardiner, C. I. 1935. Recent discoveries of reptilian remains in the Chipping Norton Limestone. *Proceedings of the Cotteswolds Naturalists' Field Club*, 25: 222.

--------- 1937. Reptile-bearing Oolite, Stow. *Report of the British Association for the Advancement of Science*, 1937 (1938): 290.

--------- 1938. Reptile-bearing Oolite, Stow. *Report of the British Association for the Advancement of Science*, 1936 (1937): 296.

Gifford, D.P. 1981. Taphonomy and paleoecology: a critical review of archaeology's sister disciplines. *In* (Schiffer, M.B.; ed.) *Advances in archaeological method and theory, vol. 4*. Academic Press, New York, 365-438.

Gilbert, A.S., Singer, B.H. & Perkins, D. 1981. Quantification experiments on computer-simulated faunal collections. *Ossa*, 8: 79-94.

Gilham, C. 1994. A fossil turtle (Reptilia: Chelonia) from the Middle Jurassic of Oxfordshire, England. *Neues Jahrbuch für Mineralogie, Geologie, und Paläontologie*, 1994: 581-596.

Gilmore, C.W. 1928. Fossil lizards of North America. *Memoirs of the National Academy of Sciences*, 22: 1-169.

Glinsky, N.L. & Bennington, J.B. 1994. Estimating numbers of whole individuals from collections of body parts: a taphonomic limitation of the palaeontological record. *Paleobiology*, 20: 245-258.

Glob, P.V. 1969. *The bog people: iron age man preserved*. Faber and Faber, London.

Gradzinski, R., Radomski, A. & Unrug, R. 1959. Preliminery results of sedimentalogical investigations in the Silesian Coal Basin. *Bulletin de l'academie Polonaise des sciences. Serie des sciences chimique, geologique et geographie*, 7: 433-438.

Grayson, D.K. 1978. Reconstructing mammalian communities: a discussion of Shotwell's method of palaeoecological analysis. *Paleobiology*, 4: 77-81.

--------- 1984. *Quantitative zooarchaeology: topics in the analysis of archaeological faunas*. Academic Press, Orlando.

Green-Kelley, R. & Mackney, D. 1970. Preferred orientation of clays in soils: the effect of drying and wetting. In (Osmond, D.A. & Bullock, P.; eds.) *Micromorphological techniques and applications.* Technical Monograph Number Two. Rothamsted Experimental Station, 43-51.

Griffin, J.J., Windom, H. & Goldberg, E.D. 1968. Distribution of clay minerals in the world ocean. *Deep-Sea Reserves.* 15: 433-461.

Guthrie, R.L. & Witty, J.E. 1982. New designations for soil horizons and the new Soil Survey Manual. *Journal of the Soil Science Society of America,* 46: 443-444.

Gvirtzman, G. & Friedman, G.J. 1977. Sequence of progressive diagenesis in coral reefs. *American Association for Petroleum Geologists, Studies in Geology,* 4:357-380.

Hagan, G.M. & Logan, B.W. 1974. Development of carbonate banks and hypersaline basins, Shark Bay, Western Australia. *American Association of Petroleum Geologists Memoir*, 22: 61-139.

Hahn, G. 1971. The dentition of the Paulchoffatiidae. *Memória 17 (N.S.) dos Serviços Geológicos de Portugal.* 39pp.

Hallam, A. 1958. The concept of Jurassic axes of uplift. *Science Progress, London*, 46: 441-449.

--------- 1975. *Jurassic Environments*. Cambridge University Press.

--------- 1978. Eustatic cycles in the Jurassic. *Palaeogeography, Palaeoclimatology, Palaeoecology.* 23: 1-32.

--------- 1981. A revised sea-level curve for the early Jurassic. *Journal of the Geological Society of London.* 138: 735-743.

--------- 1984. Humid and arid zones in the Jurassic and Cretaceous. *Palaeogeography, Palaeoclimatology, Palaeoecology.* 47: 195-223.

--------- 1985. A review of Mesozoic climates. *Journal of the Geological Society of London.* 142: 433-445.

--------- 1988. A revaluation of Jurassic eustacy in the light of new data and the revised Exxon curve. In (Wilgus, C.; ed) *Sea-level Changes- an intergrated approach*. Society of Economic Paleontologists and Mineralogists, Special Publication, 42: 261-273.

--------- 1992. Jurassic. In (Duff, P.M.D., & A.J. Smith; eds) *Geology of England and Wales.* Geological Society of London Publication. 325-354.

--------- & Sellwood, B.W. 1968. Origin of the Fullers' Earth in the Mesozoic of southern England. *Nature*, 220: 1193-1195.

--------- & --------- 1976. Middle Mesozoic sedimentation in relation to tectonics in the British area. *Journal of Geology*. 84:302-321.

Hancock, N.J. & Fisher, M.J. 1981. Middle Jurassic North Sea deltas with particular reference to Yorkshire. *In* (Illing, L.V. & Hobson, G.; eds.) *Petroleum geology of the Continental Shelf of North-West Europe*. Heyden, London. 186-195.

Hanson, C.B. 1980. Fluvial taphonomic processes: models and experiments. *In* (Behrensmeyer, A.K. & Hill, A.P.; eds.) *Fossils in the making*. University of Chicago Press, Chicago, 156-181.

Haq, B.U., Hardenbol, J. & Vail, P.R. 1987. The chronology of fluctuating sea-level since the Triassic. *Science*. 235: 1156-1167.

---------, ---------, ---------. 1988. Mesozoic and Cenozoic chronostratigraphy and cycles of sea-level change. In (Wilgus, C.; ed) *Sea-level Changes- an intergrated approach*. Society of Economic Paleontologists and Mineralogists, Special Publication, 42: 71-108.

Harland, W.B., Armstrong R.L., Cox, A.V., Craig, L.E., Smith, A.G. & Smith, D.G. 1991. *A geologic time scale 1989*. Cambridge University Press, Cambridge, 263pp.

---------, Cox, A.V., Llewellyn, P.G., Pickton, C.A.G., Smith, A.G. & Walters, R. 1982. *A geologic time scale*. Cambridge University Press, Cambridge.

Harris, J.P. & Hudson, J.D. 1980. Lithostratigraphy of the Great Estuarine Group (Middle Jurassic), Inner Hebrides. *Scottish Journal of Geology*, 16: 231-250.

Harris, T.M. 1958. Forest fire in the Mesozoic. *Journal of Ecology*, 46: 447-453.

Harrison, R.S. 1977. Caliche profiles: indicators of near-surface subaerial diagenesis, Barbados, West Indies. *Canadian Petroleum Geology Bulletin*, 25: 123-173.

Haslett, S.K. 1991. *Foraminifera from the Middle Jurassic reptile-bearing deposits of Gloucestershire biostratigraphic and palaeoenvironmental analysis.* Unpublished report. University of East Anglia, 9pp.

Haude, R. 1980. Construction morphology of the stems of Pentacrinitidae, and the way of life of *Seirocrinus*. *In* (Jangoux, M; ed.) *Echinoderms: Present and Past*, Balkema, Rotterdam, 17-23.

Haynes, G. 1980a. Evidence of carnivores gnawing on Pleistocene and Recent mammalian bones. *Paleobiology*, 6: 341-351.

--------- 1980b. Prey bones and predators: potential ecologic information from analysis of bone sites. *Ossa*, 7: 75-97.

--------- 1982. Utilisation and skeletal disturbances of North American prey carcasses. *Artic*, 35: 266-281.

--------- 1988. Mass deaths and serial predation: comparative taphonomic studies of modern large mammal death sites. Journal of Archaeological Science, 15: 219-235.

--------- 1991. *Mammoths, mastodons, and elephants: biology, behaviour, and the fossil record*. Cambridge University Press, Cambridge.

Hecht, M.K. 1992. A new choristodere (Reptilia, diapsida) from the Oligocene of France : an example of the Lazarus effect. *Geobios*, 25:115-131.

Heckel, P.H. 1983. Diagenetic model for carbonate rocks in Midcontinent Pennsylvanian eustatic cyclothems. *Journal of Sedimentary Petrology*, 53: 733-759.

Hedberg, H.B. (ed) 1976. *International Stratigraphic Guide: a guide to stratigraphic classification, terminology and procedure*. Wiley, New York, 200pp.

Hemingway, J.E. 1974. Jurassic. *In* (Rayner, D.H. & Hemingway, J.E.; eds.) *The Geology and Mineral Deposits of Yorkshire*. Yorkshire Geological Society, 161-223.

--------- & Knox, R.W. O'B. 1973. Lithostratigraphical nomenclature of the Middle Jurassic of the Yorkshire Basin of north-east England. *Proceedings of the Yorkshire Geological Society*, 39: 527-535.

Heybroek, P. 1975. Structure of the Dutch part of the Central Graben in the North Sea. *In* (Woodland, A.W.; ed.) *Petroleum and the continental shelf of north west Europe: the Geology*. Applied Science Publishers, volume 1, 339-349.

Hill, A.P. 1976. On carnivore and weathering damage to bone. *Current Anthropology*, 17: 335-336.

--------- 1978. Taphonomic background to fossil man - problems in palaeoecology. *In* (Bishop, W.W.; ed.) *Geological background to fossil man*. Scottish Academic Press, Edinburgh, 87-101.

--------- 1979a. Butchery and natural disarticulation: an investigatory technique. *American Antiquity*, 44: 739-744.

--------- 1979b. Disarticulation and scattering of mammal skeletons. *Paleobiology*, 5: 261-274.

--------- 1980. Early postmortem damage to the remains of some contemporary East African mammals. *In* (Behrensmeyer, A.K. & Hill, A.P.; eds.) *Fossils in the making: Vertebrate taphonomy and palaeoecology*. University of Chicago Press.

--------- & Behrensmeyer, A.K. 1984. Disarticulation patterns of some modern East African mammals. *Paleobiology*, 10: 366-376.

--------- & --------- 1985. Natural disarticulation and bison butchery. *American Antiquity*, 50: 141-145.

--------- & Walker, A. 1972. Procedures in vertebrate taphonomy. *Journal of the Geological Society of London*, 128: 399-406.

Hine, A.C. 1977. Lily Bank, Bahamas: history of an active oolite sand shoal. *Journal of Sedimentary Petrology*, 47: 1554-1582.

--------- , Belknap, D.F., Hutton, J.G., Osking, E.B. & Evans, M.W. 1988. Recent geological history and modern sedimentary processes along an incipient, low-

energy, epicontinental-sea coastline: northwest Florida. *Journal of Sedimentary Petrology*, 58:567-579.

Holloway, S. 1983. The shell-detrital calcirudites of the Forest Marble Formation (Bathonian) of southwest England. *Proceedings of the Geologists' Association*, 94:259-266.

Holmes, A. 1978. *Principles of Physical Geology*. Van Nostrand Reinhold (U.K.) Co. Ltd., Wokingham.

Holzhey, C.S., Yeck, R.D. & Nettleton, W.D. 1974. Microfabric of some argillic horizons in udic, xeric and torric soil environments of the United States. *In* (Rutherford, G.K.; ed.) *Soil Microscopy*. Limestone, Kingston, 747-760.

Holtzman, R.C. 1979. Maximum likelihood estimation of fossil assemblage composition. *Paleobiology*, 5: 77-89.

Horton, A. 1977. The Age of the Middle Jurassic "white sands" of North Oxfordshire. *Proceedings of the Geologists' Association*, 88: 147-162.

--------- , Poole, E.G., Williams, B.J., Illing, V.C. & Hobson, G.D. 1987. *Geology around Chipping Norton*. Memoir of the British Geological Survey, 218. HMSO, London.

Howitt, F., Aston, E.R. & Jacque, M. 1975. The occurrence of Jurassic volcanics in the North Sea. *In* (Woodland, A.W.; ed.) *Petroleum and the Continental Shelf of North West Europe. Volume 1: Geology*. Applied Science, Barking, 379-387.

Hudleston, W.H. 1878. Excursion to Chipping Norton. *Proceedings of the Geologists' Association*, 10: 166-172.

Hudson, J.D. 1962. The stratigraphy of the Great Estuarine Series (Middle Jurassic) of the Inner Hebrides. *Transactions of the Edinburgh Geological Soc*iety, 19: 139-165.

--------- 1963. The recognition of salinity-controlled mollusc assemblages in the Great Estuarine Series (Middle Jurassic) of the Inner Hebrides. *Palaeontology*. 6: 318-326.

--------- 1964. The petrology of the sandstones of the Great Estuarine Series, and the Jurassic palaeogeography of Scotland. *Proceedings of the Geologists' Association*, 75: 499-527.

--------- 1966. Hugh Miller's Reptile Bed and the *Mytilus* Shales, Middle Jurassic, Isle of Eigg, Scotland. *Scottish Journal of Geology*, 2: 265-281.

--------- 1983. Mesozoic sedimentation and sedimentary rocks in the Inner Hebrides. *Proceedings of the Royal Society of Edinburgh*. 83B:47-63.

--------- & Morton, N. 1969. *International field symposium on the British Jurassic, Part IV(Scotland)* Keele University Press, 47pp.

--------- & Palmer, T.J., 1976. A euryhaline oyster from the Middle Jurassic and the origin of the true oysters. *Palaeontology*. 19: 79-93.

Huene, F. von. 1906. Über das Hinterhaupt von *Megalosaurus bucklandi* aus Stonesfield. *Neues Jahrbuch für Mineralogie, Geologie, und Paläontologie*, 1906: 1-12.

--------- 1910. Über den ältesten Rest von *Omosaurus (Dacentrurus)* im englischen Dogger. *Neues Jahrbuch für Mineralogie, Geologie, und Paläontologie*, 1910: 75-78.

Hull, E. 1855. On the physical geography of the Cotteswold Hills. *Quarterly Journal of the Geological Society of London*, 11: 477-494.

--------- 1859. *The geology of the country around Woodstock, Oxfordshire*. Memoirs of the Geological Survey of the Great Britain, 45. HMSO, London. 30pp.

--------- 1860. On the Blenheim iron-ore and the thickness of the formations below the Great Oolite at Stonesfield, Oxfordshire. *The Geologist*, 3: 303-305; and *Report of the British Association for the Advancement of Science*, 1860 (1861): 81-83.

Humboldt, F.W.H.A. von 1795. *Ueber die unterirdischen Gasarten und die Mittel ihren Nachtheil zu vermindern. Ein Beytrag zur Physik der practischen Bergbaukunde*. F. Weigwag, Braunschweig, 384pp.

Hunt, R.M., Jr. 1978. Depositional setting of a Miocene mammal assemblage, Sioux County, Nebraska (U.S.A.). *Palaeogeography, Palaeoclimatology, Palaeoecology*, 24: 1-52.

Hurst, A. 1985. The implications of clay mineralogy to palaeoclimate and provenance during the Jurassic of Scotland. *Scottish Journal of Geology*, 21: 143-160.

Hutchinson, G.E. 1957. *A treatise on limnology, Volume 2. Introduction to lake biology and limnoplankton*. John Wiley, New York, 1115pp.

Huxley, T.H. 1859. On Rhamphorhynchus bucklandi, a pterosaurian from the Stonesfield Slate. *Quarterly Journal of the Geological Society of London*, 15: 678-680.

Ivanovich, M. & Ireland, P. 1984. Measurement of uranium series disequilibrium in the case-hardened Aymaman Limestone of Puerto Rico. *Zeitschrift für Geomorphologie*, 28: 305-319.

Jeans, C. V., Merriman, R.J. & Mitchell, J.G. 1977. Origin of Middle Jurassic and Lower Cretaceous Fuller's Earths in England. *Clay Minerals*, 12: 11-44.

Jennings, J.N. 1971. *Karst*. M.I.T. Press, London.

--------- 1975. Doline morphometry as a morphological tool: New Zealand examples. *New Zealand Geographer*, 31: 6-28.

--------- 1985. *Karst geomorphology*. Blackwell, Oxford.

Johnson, E. 1985. Current developments in bone technology. *In* (LeMoine, G.M. & MacEachern, A.S.; eds.) *Carnivores, human scavengers and predators: a question of bone technology*. University of Calgary Archaeological Association, Calgary, 55-93.

Johnson, R.G. 1957. Experiments on the burial of shells. *Journal of Geology*, 65: 527-535.

--------- 1960. Models and methods for analysis of the formation of fossil assemblages. *Bulletin of the Geological Society of America*, 71: 1075-1086.

--------- 1962. Mode of formation of marine fossil assemblages of the Pleistocene Millerton Formation of California. *Bulletin of the Geological Society of America*, 73: 113-130.

Jones, T.P. 1993. New morphological and chemical evidence for a wildfire origin for fusain from comparisons with modern charcoal. *Special Papers in Palaeontology*, 49: 113-123.

--------- , Scott, A.C. & Mattey, D.P. 1993. Investigations of "fusain transition fossils" from the Lower Carboniferous: comparisons with modern partially charred wood. *International Journal of Coal Geology*, 22: 37-59.

Kahle, C.F. 1977. Origin of subaerial Holocene calcareous crusts: role of algae, fungi and sparmicritisation. *Sedimentology*, 24: 413-436.

Kellaway, G.A. & Welch, F.B.A. 1948. *Bristol and Gloucester District.* NERC Institute of Geological Sciences. HMSO, London.

Kermack, K.A. 1988. British Mesozoic mammal sites. *Special papers in Palaeontology*, 40:85-93.

--------- , Lee, A.J., Lees, P.M., and Mussett, F. 1987. A new docodont from the Forest Marble. *Zoological Journal of the Linnean Society*, 89:1-39.

Kidwell, S.M. 1991. The stratigraphy of shell concentrations. *In* (Allison, P.A. & Briggs, D.E.G.; eds.) *Taphonomy: releasing the data locked in the fossil record.* Volume 9 of Topics in Geobiology. Plenum Press, New York, 212-279.

--------- & Aigner, T. 1983. Sedimentary dynamics of complex shell beds: implications for ecologic and evolutionary patterns. *In* (Bayer, U., & Seilacher, A.; eds.) *Sedimentary and evolutionary cycles*. Springer-Verlag, Berlin, 382-395.

--------- & Behrensmeyer, A.K. 1988. Ecological and evolutionary implications of taphonomic processes. *Palaeogeography, Palaeoclimatology, Palaeoecology, Special Issue*, 63: 1-33.

--------- & Bosence, D.W.J. 1991. Taphonomy and time-averaging of marine shelly faunas. *In* (Allison, P.A. & Briggs, D.E.G.; eds.) *Taphonomy: releasing the data locked in the fossil record.* Volume 9 of Topics in Geobiology. Plenum Press, New York, 115-209.

--------- , Fürsich, F.T. & Aigner, T. 1986. Conceptual framework for the analysis and classification of fossil concentrations. *Palaios*, 1: 228-238.

--------- & Jablonski, D. 1983. Taphonomic feedback: ecological consequences of shell accumulation. *In* (Tevesz, M.J.S. & McCall, P.L.; eds.) *Biotic interactions in recent and fossil benthic communities*. Plenum Press, New York, 195-248.

Kinsman, D.J.J. 1976. Evaporites: relative humidity control of primary mineral facies. *Journal of Sedimentary Petrology*. 46: 273-279.

Klein, R.G. & Cruz-Uribe, K. 1984. *The analysis of animal bones from archaeological sites*. University of Chicago, Chicago.

Klute, A. 1986. *Methods of soil analysis. Part 1. Physical and mineralogical methods.* American Society of Agronomy, Madison.

Komarek, E.V. 1972. Ancient fires. *Proceedings of the Annual Tall Timbers Fire Ecology Conference*, 12: 219-240.

Korth, W.W. 1979. Taphonomy of microvertebrate fossil assemblages. *Annals of the Carnegie Museum*, 48: 235-285.

Kreutzer, L. Bone orientation data for the Clovis Age Bed at the Lubbock Lake Landmark. *Current Research in the Pleistocene*, 4: 79-80.

Kühne, W.G. 1946. The geology of the fissure infilling "Holwell 2"; the age determination of the mammalian teeth therein; and a report on the technique employed when collecting the teeth of *Eozostrodon* and Microclepidae. *Proceedings of the Zoological Society of London*, 116: 729-733.

--------- 1956. *The Liassic therapsid* Oligokyphus. British Museum (Natural History): London 145pp.

Land, L.S. 1971. Phreatic versus meteoric diagenesis of limestones: evidence from a fossil water table in Bermuda. *In* (Bricker, O.D.; ed.) *Carbonate Cements*. John Hopkins Press, Baltimore, 133-137.

Lawrence, D.R. 1968. Taphonomy. *In* (Fairbridge, R. and Jablonski, D., eds.) *Encyclopedia of Paleontology, vol. 7.* Dowden, Hutchinson and Ross, New York, 793-799.

Leeder, M.R. & Nami, M. Sedimentary models for the non-marine Scalby Formation (Middle Jurassic) and evidence for Late Bajocian/Bathonian uplift of the Yorkshire Basin. *Proceedings of the Yorkshire Geological Society*, 42: 461-482.

Linslay, P.N., Potter, H.C., McNab, G. & Racher, D. 1980. The Beatrice Field, Inner Moray Firth, U.K. North Sea. *Ninian Field, U.K. Sector, North Sea. American Association of Petroleum Geologists*, Memoir 30: 117-129.

Livera, S.E. & Leeder, M.R. 1981. The Middle Jurassic Ravenscar Group ('Deltaic Series') of Yorkshire. *Proceedings of the Geologists' Association*, 92: 421-250.

Lhuyd, E. 1699 *Eduardi Luidii apud oxonienses cimeliarchae ashmoleani lithophylacii britannici ichnographia*. London, 145pp.

Lockley, M. 1991. *Tracking dinosaurs: A new look at an ancient world.* Cambridge University Press, Cambridge.

Longman, M.W. 1980. Carbonate diagenetic textures from near surface diagenetic environments. *Bulletin of American Association of Petroleum Geologists'*, 64: 461-487.

Lott, G.K. & Humphreys, B. 1992. The stratigraphy and petrology of Middle Jurassic (Ravenscar Group) sediments in cored boreholes from the north Yorkshire coast. *Proceedings of the Yorkshire Geological Society*, 49: 23-40.

Lucas, J. & Prevot, L.E. 1991. Phosphates and fossil preservation. In (Allison, P.A. & Briggs, D.E.G.; eds.) *Taphonomy: releasing the data locked in the fossil record. Volume 9 of Topics in Geobiology*. Plenum Press, New York, 389-405.

Lyman, R.L. 1994. *Vertebrate taphonomy. Cambridge Manuals in Archaeology*. Cambridge University Press, Cambridge.

--------- & Fox, G.L. 1989. A critical evaluation of bone weathering as an indication of bone assemblage formation. *Journal of Archaeological Science*, 16: 293-317.

Maas, M.C. 1984. Taphonomy of a late Eocene microvertebrate locality, Wind River Basin, Wyoming (U.S.A.). *Palaeogeography, Palaeoclimatology, Palaeoecology*, 52: 123-142.

Mack, G.H., James, W.C. & Monger, H.C. 1993. Classification of paleosols. *Geological Society of America Bulletin*, 105: 129-136.

Maisey, J.G. 1977. The fossil selachian fishes *Palaeospinax* Egerton 1872 and *Nemacanthus* Agassiz 1837. *Zoological Journal of the Linnean Society*, 60, 259-73.

Manabe, S. & Wetherland, R.T. 1980. On the distribution of climatic changes resulting from an increase in CO2 content in the atmosphere. *Journal of Atmospheric Science*, 37: 99-118.

Mantell, G. A. 1822. *The Fossils of the South Downs; or, Illustrations of the Geology of Sussex*. L. Relfe, London, 327 pp.

Marshall, L.G. 1989. Bone modification and "the laws of burial". *In* (Bonnichsen, R. & Sorg, M.H.; eds.) *Bone modification*. University of Maine Center for the Study of the First Americans, Orono, 7-24.

Martill, D.M. 1986. The stratigraphic distribution of fossil vertebrates in the Oxford Clay of England. *Mercian Geologist*, 10: 161-186.

--------- 1989. Fungal borings in neoselachian teeth from the Lower Oxford Clay of Peterborough. *Mercian Geologist*, 12: 1-4.

--------- 1991. Bones as stones: the contribution of vertebrate remains to the lithologic record. *In* (Donovan, S.K.; ed.) *The Processes of Fossilisation*. Belhaven Press, London, 270-293.

--------- & Hudson, J.D. 1991 (eds.). *Fossils of the Oxford Clay. Palaeontological Association field guide to fossils: 4*. The Palaeontological Association, London, 286pp.

Martin, A.J. 1967. Bathonian sedimentation in southern England. *Proceedings of the Geological Society of London*, 78: 473-488.

Massare, J.A. 1987. Tooth morphology and prey preference of Mesozoic marine reptiles. *Journal of Vertebrate Paleontology*, 7: 121-137.

Masson, D.G. & Miles, P.R. 1986. Development and hydrocarbon potential of Mesozoic sedimentary basins around margins of North Atlantic. *Bulletin of the American Association of Petroleum Geologists*, 70: 721-729.

Matley, D.K. 1941. The coprolites of Pijdura, Central Provinces. *Records of the Geological Survey of India*, 74: 535-547.

Mauberge, P.-L. 1964. Resolutions du Colloque. *In* (Mauberge, P.L.; ed.) *Colloque du Jurassique a Luxembourg 1962*. Luxembourg, 77-80.

Mayhew, D.F. 1977. Avian predators as accumulators of fossil mammal material. *Boreas*, 6: 25-31.

McKerrow, W. S. & Baden-Powell, D. F. W. 1953. Easter field meeting, 1952: The Jurassic rocks of Oxfordshire and their superficial deposits. *Proceedings of the Geologists' Association*, 64: 88-98.

--------- & Baker, S. 1988. Field meeting to Charlbury and Stonesfield, Oxfordshire. *Proceedings of the Geologists' Association*, 99: 61-65.

--------- Johnson, R. T. & Jakobson, M. E. 1969. Palaeoecological studies in the Great Oolite at Kirtlington, Oxfordshire. *Palaeontology*, 12: 56-83.

--------- & Kennedy, W.J. 1973. *The Oxford District.* Geological Association Guide No. 3, 1-46.

Megnien, C. & Megnien, F. (eds.) 1980. Synthese geologique du bassin de Paris. *Memoires du Bureau de Recherches Geologiques et Minieres*, 101-103.

Mellet, J.S. 1974. Scatological origin of microvertebrate fossil accumulations. *Science*, 185: 349-350.

Metcalf, S.J. 1993. The palaeoenvironment of a new British dinosaur locality from the lower Bathonian (Middle Jurassic). *In* (Buffetaut, E. & Mazin, J.-M.; eds) *Dinosaurs and oth fossil reptiles from Europe. Proceedings of the Second Georges Cuvier Symposium.* Revue de Paleobiologie, Volume Special, 7: 125-149.

--------- Vaughan, R.F., Benton, M.J., Cole, J., Simms M.J. & Dartnall, D.L. 1992. A new Bathonian (Middle Jurassic) microvertebrate site, within the Chipping Norton Limestone Formation at Hornsleasow Quarry, Gloucestershire. *Proceedings of the Geologists' Association*, 103: 321-342.

--------- & Walker, R.J. 1994. A new Bathonian microvertebrate locality in the English Midlands. *In* (Fraser, N.C. and H.-D. Sues; eds) *In the shadow of the dinosaurs: early Mesozoic tetrapods*. Cambridge University Press, New York, 322-331.

--------- & Underwood, C.J. 1995 (in prep.) Depositional environment of the Charlbury and Taynton Limestone formations (Bathonian, Middle Jurassic) of the northern Cotswolds.

Miller H. 1858 *The Cruise of the Betsey.* Edinburgh.

Milner, A.R. 1993. Amphibian-Grade Tetrapoda. In (Benton, M.J.; ed.) *The Fossil Record 2.* Chapman & Hall, London, p. 665-680.

--------- & Evans, S.E. 1991. The upper Jurassic diapsid *Lisboasaurus estesi* - a maniraptoran theropod. *Palaeontology*, 34: 503-513.

Molnar, R.E. 1990. Problematic Theropoda: "carnosaurs". *In* (Weishampel, D. B., Dodson, P., & OsmÓlska, H.; eds.) *The Dinosauria*. University of California Press, Berkeley, 169-209.

Morton, N. 1965. The Bearraraig Sandstone Series (Middle Jurassic) of Skye and Raasay. *Scottish Journal of Geology*, 1: 189-216.

--------- 1971. The definition of standard Jurassic stages. *In* (Morton, N; ed.) *Colloque du Jurassique a Luxembourg 1967. Memoires du Bureau de Recherches Geologiques et Minieres.* 75:83-93.

--------- 1987. Jurassic subsidence history in the Hebrides, north-west Scotland. *Marine and Petroleum Geology*, 4: 226-242.

Mudge, D.C. 1978. Stratigraphy and sedimentation of the Lower Inferior Oolite of the Cotswolds. *Journal of the Geological Society of London,* 135: 611-627.

Muller, H.G. & Hoffmeister, J.E. 1968. Subaerial laminated crusts of the Florida Keys. *Geological Society of America Bulletin*, 79: 183-192.

Myers, T., Voorhies, M.R. & Corner, R.G. 1980. Spiral fractures and bone psuedotools at palaeontological sites. *American Antiquities*, 45: 483-489.

Nami, M. 1976. An exhumed Jurassic meander belt from Yorkshire, England. *Geological Magazine*, 133: 47-52.

--------- & Leeder, M.R. 1978. Changing channel morphology and magnitude in the Scalby Formation (M. Jurassic) of Yorkshire, England. *In* (Miall, A.D.; ed.) *Fluvial Sedimentology.* Canadian Society of Petroleum geologists Memoir, 5.

Neuwmann, A.C. & Land, L.S. 1967. Lime mud deposition and calcareous algae in the Bight of Abaco, Bahamas. *Journal of Sedimentary Petrology*, 45: 763-786.

Newesley, H. 1989. Fossil bone apatite. *Applied Geochemistry*, 4: 233-246.

Newton, E.F. 1937. The petrography of some English Fullers Earth and the rocks associated with them. *Proceedings of the Geologists' Association*, 48: 175-197.

Odling, M. 1913. The Bathonian rocks of the Oxford District. *Quarterly Journal of the Geological Society of London*, 69:484-513.

Olsen, E.C. & Shipman, P. 1988. Surface modification on bone: Trampling versus butchery. *Journal of Archaeological Research*, 15: 535-553.

Olson, E.C. 1952. The evolution of Permian vertebrate chronofauna. *Evolution*, 6: 181-196.

--------- 1958. Fauna of the Vale of Choza: 14. Summary, review, and integration of the geology and the faunas. *Fieldiana Geology*, 10: 397-448.

Oppel, A. 1856. Ueber einige Cephalopoden der Juraformtion Wurttembergs. *Wurttenburg Naturwissenschaftlichen Jahresheften*, 12pp.

Owen, R. 1838a. On the jaws of the *Thylacotherium prevostii* (Valanciennes) from Stonesfield. *Proceedings of the Geological Society of London*, 3: 5-9.

--------- 1838b. A paper on the 'Phascolotherium', being the second part of the 'Description of the remains of marsupial Mammalia from the Stonesfield Slate'. *Ibid* : 17-21.

--------- 1842. Observations on the fossils representing the *Thylacotherium prevostii* (Valanciennes), with reference to the doubts of its mammalian and marsupial nature recently promulgated; and on the *Phascolotherium bucklandi*. *Transactions of the Geological Society of London*, 6: 45-65.

--------- 1857. On the affinities of the *Stereognathus ooliticus* (Charlesworth) a mammal from the Oolitic Slate of Stonesfield. *Quarterly Journal of the Geological Society of London*, 13: 1-11.

--------- 1871. Monograph of the fossil Mammalia of the Mesozoic formations. *Palaeontographical Society [Monograph]*, 6: 115pp.

Palmer, T.J. 1972. Field meeting in the Great Oolite of Oxfordshire. *Proceedings of the Geologists' Association*, 84: 53-64.

--------- 1979. The Hampen Marly and White Limestone Formations: Florida-type carbonate lagoons in the Jurassic of central England. *Palaeontology*, 22: 189-228.

--------- & Jenkyns, H.C. 1975. A carbonate island barrier from the Great Oolite (Middle Jurassic) of central England. *Sedimentology*, 22: 125-135.

Parfenova, E.I. & Yaribova, E.A. 1962. *Mineralogical investigations in soil science*. Moscow Academy of Science.

Parrish, J.T., Ziegler, A.M. & Scotese, C.R. 1982. Rainfall patterns and distribution of coals and evaporites in the Mesozoic and Cenozoic. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 40: 67-102.

Parsons, C.F. 1976a. Ammonite evidence for dating some Inferior oolite sections in the north Cotswolds. *Proceedings of the Geologists Association*. 87:45-63.

--------- 1976b. A statigraphic revision of the *humphriesianum-subfurcatum* Zone rocks (Bajocian Stage, Middle Jurassic) of southern England. *Newsletters in Stratigraphy*. 5:114-142.

Patterson, C. 1966. British Wealden sharks. *Bulletin of the British Museum (Natural History), Geology Series*, 11: 281-350.

Perkins, R.D. 1977. Depositional framework of Pleistocene rocks in South Florida. *Geological Society of America Memoir*, 147: 131-198.

Phillips, J. 1871. *Geology of Oxford and the valley of the Thames*. Clarendon, Oxford., 523pp.

Phizackerley, P. H. 1951. A revision of the Teleosauridae in the Oxford University Museum (Natural History). *Annals and Magazine of Natural History, Series 12*, 4: 1169-1192.

Pitman, W.C. 1978. Relationship between eustacy and stratigraphic sequences at passive margins. *Geological Society of America Bulletin*. 89:1389-1403.

--------- & Talwani, M. 1972. Seafloor spreading in the North Atlantic. *Geological Society of America Bulletin*. 83:619-646.

Platt, J. 1758. An account of the fossile thigh-bone of a large animal, dug up at Stonesfield, near Woodstock, in Oxfordshire. *Philosophical Transactions of the Royal Society*, 50: 524-527.

Plot, R. 1677 *The natural history of Oxfordshire, being an essay toward the natural history of England*. Theatre, Oxford, 358pp.

Potter, P.E. & Pettijohn, F.J. 1977. *Palaeocurrents and basin analysis*. Springer-Verlag, New York.

Potts, R. 1986. Temporal span of bone accumulations at Olduvai Gorge and implications for early hominid foraging behaviour. *Paleobiology*, 12: 25-31.

Pratt, A.E. 1989. Taphonomy of the microvertebrate fauna from the early Miocene Thomas Farm locality, Florida (U.S.A.). *Palaeogeography, Palaeoclimatology, Palaeoecology*, 76: 125-151.

Prevost, L.C. 1825. Observations sur les schistes calcaires oolitiques de Stonesfield en Angleterre. *Annals de la Science naturelle*, 4 : 389-416.

Prevot, L. & Lucas, J. 1990. Phosphate. *In* (Briggs, D.E.G. & Crowther, P.R.; eds.) *Palaeobiology: a synthesis*. Blackwell, Oxford, 256-257.

Purser, B.H. 1969. Syn-sedimentary marine lithification of Middle Jurassic limestones in the Paris Basin. *Sedimentology*, 12: 205-230.

Raup, D.M. 1976. Species diversity in the Phanerozoic: a tabulation. *Paleobiology*, 2: 279-288.

--------- 1979. Biases in the fossil record of species and genera. *Bulletin of the Carnegie Museum of Natural History*, 13: 85-91.

Read, H.H. & Watson, J. 1975. *Introduction to Geology Volume 2, Earth History*. Macmillan, London.

Read, J.F. 1974. Calcrete deposits and Quaternary sediments, Edel Province, Western Australia. *American Association of Petroleum Geologists Memoir*, 22: 250-282.

Retallack, G.J. 1976. Triassic paleosols in the upper Narrabeen Group of New South Wales. Part 1, features of the paleosols. *Journal of the Geological Society of Australia*, 23: 383-399.

--------- 1977. Triassic paleosols in the upper Narrabeen Group of New South Wales. Part 2, classification and reconstruction. *Journal of the Geological Society of Australia*, 24: 19-35.

--------- 1984. Completeness of the rock and fossil record; some estimates using fossil soils. *Paleobiology*, 10: 59-78.

--------- 1985. Fossil soils as grounds for interpreting the advent of large plants and animals on land. *Philosophical Transactions of the Royal Society of London*, B309: 105-142.

--------- 1986. The fossil record of soils. *In* (Wright, V.P.; ed.) *Paleosols: their recognition and interpretation*. Oxford, Blackwell, 1-57.

--------- 1990. *Soils of the past - an introduction to paleopedology*. Unwin Hyman, Boston, 520pp.

--------- 1991. Untangling the effects of burial alteration and ancient soil formation. *Annual Review of Earth and Planetary Sciences*, 19: 193-206.

--------- 1992. What to call early plant formations on land. *Palaios*, 7: 508-520.

--------- & Dilcher, D.L. 1981. A coastal hypothesis for the dispersal and rise to dominance of flowering plants. *In* (Niklas, K.J.; ed.) *Paleobotany, palaeoecology and evolution*, 2: 27-77. Praeger, New York.

Reynolds, S. H. 1939. On a collection of reptilian bones from the Oolite of Stow-on-the-Wold, Gloucestershire. *Geological magazine*, 76: 193-214.

Richards, B.N. 1987. *The microbiology of terrestrial ecosystems*. Longman Scientific and Techinical, Harlow.

Richards, P.C., Brown, S., Dean, J.M. & Anderton, R. 1988. A new palaeogeographic reconstruction for the Middle Jurassic of the northern North Sea. *Journal of the Geological Society, London*, 145: 883-886.

Richardson, L. 1911. The Inferior Oolite and contiguous deposits of the Chipping Norton district. *Proceedings of the Cotteswold Naturalists Field Club*, 17: 203-206.

--------- 1929. *The country around Moreton-in-the-Marsh*. Memoir of the Geological Survey of Great Britain, 217. HMSO, London. 162pp.

--------- , Arkell, W. J., and Dines, H. G. (1946) *Geology of the country around Witney*. Memoir of the Geological Survey of Great Britain, 236. HMSO, London, 150pp.

Richter, R. 1928. *Aktuopaläontologie und paläobiologie, eine abgrenzung*. Senkenbergiana 19.

Riding, J.B. & Wright, J.K. 1989. Palynostratigraphy of the Scalby Formation (Middle Jurassic) of the Cleveland Basin, north-east Yorkshire. *Proceedings of the Yorkshire Geological Society*, 47: 349-354.

Riding, R. & Wright, V.P. 1981. Paleosols and tidal-flat/lagoonal sequences on a Carboniferous carbonate shelf: sedimentary associations of triple disconformities. *Journal of Sedimentary Petrology*, 51:1323-1339.

Robinson, D. & Wright, V.P. 1987. Ordered illite/smectite and kaolinite/smectite as possible primary minerals in a Lower Carboniferous palaeosol sequence, South Wales? *Clay Minerals*, 22: 109-118.

Robinson, P.L. 1971. A problem of faunal replacement on Permo-Triassic continents. *Palaeontology*. 14: 131-153.

--------- 1973. Palaeoclimatology and continental drift. *In* (Tarling, D.H. & Runcorn, S.K.; eds.) *Implications of Continental Drift to the Earth Sciences*. Academic Press, London. Volume 1: 449-476.

Rock-Color Chart Committee 1991. *Rock color chart, with genuine Munsell^{R} color chips*. Geological Society of America.

Rolfe, W.D.I. & Brett, D.W. 1969. Fossilisation processes. *In* (Eglington, G. & Murphy, M.T.J.; eds.) *Organic geochemistry: methods and results*. Springer-Verlag, Berlin, 214-244.

Romer, A. S. 1956. *Osteology of the reptiles*. University of Chicago Press, Chicago, 772 pp.

Romer, A. S. 1966. *Vertebrate paleontology*. 3rd edn. University of Chicago Press, Chicago, 468 pp.

Romano, M. & Spears, D.A. 1991. Bentonites from the Horton Formation (Upper Silurian) of Ribblesdale, Yorkshire. *Proceedings of the Yorkshire Geological Society*, 48: 277-285.

Ronov, A.B. & Khain, V.Y. 1962. Jurassic formations of the world. *Sovetsk. Geol.* 1: 9-34.

Ruhle, R.V., Cady, J.G. & Gomez, R.Z. 1961 Paleosols of Bermuda. *Geological Society of America Bulletin*, 72: 1121-1142.

Russell, E.W. 1988. *Soil conditions and plant growth*. Longman, London.

Russell, R.S. 1977. *Plant root systems: their function and interaction with the soil.* McGraw-Hill, London.

Sargeant, W.A.S. 1970. Fossil footprints from the Middle Triassic of Nottinghamshire and the Middle Jurassic of Yorkshire. *Mercian Geologist*, 3:269-282.

Sargeant, W.A.S. 1974. A history and bibliography of the study of fossil vertebrate footprints in the British Isles. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 16: 263-378.

Savage, R. J. G. 1963. The Witts collection of Stonesfield Slate fossils. *Proceedings of the Cotteswold Naturalists' Field Club*, 33: 177-182.

--------- 1984. Mid-Jurassic mammals from Scotland. 211-213. *In* (Reif, W.-E. and Westphal, F.; eds.) *Third Symposium on Mesozoic Terrestrial Ecosystems.* Attempto, Tübingen.

Schopf, J.M. 1975. Modes of fossil preservation. *Review of Palaeobotany and Palynology*, 20: 27-53.

Scott, A.C. 1990. Anatomical preservation of fossil plants. *In* (Briggs, D.E.G. & Crowther, P.R.; eds.) *Palaeobiology: a synthesis*. Blackwell, Oxford, 263-266.

--------- & Jones, T.P. 1993. The nature and influence of fire in Carboniferous ecosystems. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 106: 91-112.

Seilacher, A. 1977. SFB 53 'Palökologie', Tübingen. *Lethaia*, 10: 200.

Sellwood, B.W. 1978. Jurassic. *In* (McKerrow, W.S.) *The ecology of fossils.* Duckworth, London, 204-279.

--------- 1986. Shallow-marine carbonate environments. *In* (Reading, H.G.; ed.) *Sedimentary environments and facies*. Blackwell Scientific Publications, Oxford.

--------- & Hallam, A. 1974. Bathonian volcanicity and North Sea rifting. *Nature*, 252: 27.

--------- & Jenkyns, H.C. 1975. Basins and swells and the evolution of an epeiric sea (Pliensbachian-Bajocian of Great Britain). *Journal of the Geological Society, London*, 131: 373-388.

--------- & McKerrow, W.S. 1974. Depositional environments in the lower part of the Great Oolite Group of Oxfordshire and north Gloucestershire. *Proceedings of the Geologists Association.*, 85: 190-210.

Sepkoski, J.J. 1976. Species diversity in the Phanerozoic: species area effects. *Paleobiology*, 2: 298-303.

Sereno, P. C. & Dong Z.-M. 1992. The skull of the basal stegosaur *Huayangosaurus taibii* and a cladistic diagnosis of Stegosauria. *Journal of Vertebrate Paleontology*, 12: 318-343.

Shaw, A.B. 1964. *Time in stratigraphy*. McGraw-Hill, New York.

Shearman, D.J., Tyman, J. & Kandkarimi, M. 1970. The genesis and diagenesis of oolites. *Proceedings of the Geologists' Association*, 81: 1905-1914.

Shinn, E.A. 1983. Birdseyes, fenestrae, shrinkage pores and loferites: a re-evaluation. *Journal of Sedimentary Petrology*, 53:619-629.

--------- & Lidz, B.H. 1988. Blackened limestone pebbles: fire at subaerial unconformities. *In* (James, N.P. & Choquette, P.W.; eds.) *Palaeokarst.* Springer-Verlag, New York, 117-131.

Shipman, P. 1981. *Life history of a fossil, an introduction to taphonomy and palaeoecology*. Harvard University Press, Massachusetts.

--------- , Bosler, W. & Davis, K.L. 1981. Butchering of giant geladas at an Acheulian site. *Current Anthropology*, 22: 257-268.

--------- , Fisher, D.C. & Rose, J.J. 1984. Mastodon butchery: microscopic evidence of carcass processing and bone tool use. *Paleobiology*, 10: 358-365.

--------- & Rose, J.J. 1983. Evidence of butchery activities at Torralba and Ambrona: an evaluation using microscopic techniques. *Journal of Archaeological Science*, 10: 465-474.

Shotwell, J.A. 1955. An approach to the palaeoecology of mammals. *Ecology*, 36: 327-337.

Simms, M.J. 1986. Contrasting lifestyles in Lower Jurassic crinoids : a comparison of benthic and pseudopelagic isocrinida. *Palaeontology*, 29: 475-493.

Simpson, G.G. 1928. *A catalogue of the Mesozoic Mammalia in the Geological Department of the British Museum*. British Museum (Natural History), London.

Singer, A. 1980. The palaeoclimatic interpretation of clay minerals in soils and weathering profiles. *Earth Science Reviews* , 15: 303-326.

Skerl, J.G.A. 1925. The petrography of certain Jurassic rocks from the North Cotswolds. *Proceedings of the Cotteswolds Naturalists' Field Club*, 22: 281-286.

Smith, A.G. 1971. Alpine deformation and oceanic areas of the Tethys, Mediterranean, and Atlantic. *Bulletin of the Geological Society of America*, 82: 2039-2070.

--------- & Briden, J.C. 1977. *Mesozoic and Cenozoic Paleocontinental Maps*. Cambridge University Press.

--------- , --------- & Drewry, G.E. 1973. Phanerozoic world maps. *Special Papers in Palaeontology*, 12:1-39.

Soil Survey Staff. 1975. *Soil taxonomy: United States Department of Agriculture (U.D.S.A.) handbook 436*. Government Printing Office, Washington.

Spicer, R.A. 1991. Plant taphonomic processes. In (Allison, P.A. & Briggs, D.E.G.; eds.) *Taphonomy: releasing the data locked in the fossil record. Volume 9 of Topics in Geobiology*. Plenum Press, New York, 71-113.

Srodon, J. & Eberl, D.D. 1984. Illite. In (Bailey, S.W.; ed.) *Micas. Reviews in Mineralogy, 13*. Mineralogical Society of America, 495-538.

Stace, H.C.T., Hubble, G.D., Brewer, R., Northcote, K.H., Sleeman, J.R., Mulcahy, M.J. & Hallsworth, E.G. 1968. *A handbook of Australian soils*. Rellim Technical Publications, Glenside.

Steel, R. 1985. *Sharks of the World*. Blandford Press, London. 192pp.

Storrs, G. W. & Gower, D. J. 1993. *Pachystropheus (Rysosteus)* from the Rhaetic Bone-beds of southern England, the world's oldest choristodere. *Journal of the Geological Society of London*, 150: 1103-1107.

Strahan, A. 1898. The Geology of the Isle of Purbeck and Weymouth (Sheets 341-3). *Memoirs of the Geological Survey of the United Kingdom*, HMSO, London, 196pp.

Strasser, A. 1984. Black pebble occurrence and genesis in Holocene carbonate sediments (Florida Keys, Bahamas, and Tunisia). *Journal of Sedimentary Petrology*, 54:1097-1109.

--------- & Davaud, E. 1983 Black pebbles of the purbeckian (Swiss and French Jura): lithology, geochemistry and origin. *Eclogae geol. Helv.*, 76:551-580.

Stucky, R.K. & McKenna, M.C. 1993. Mammalia. *In* (Benton, M.J.; ed.) *The Fossil Record 2*. Chapman & Hall, London, p. 739-774.

Sun, S.Q. 1990. Facies-related diagenesis in a cyclic shallow marine sequence: the Corallian Group (Upper Jurassic) of the Dorset Coast, Southern England. *Journal of Sedimentary Petrology*, 60:42-52.

Sweeting, M.M. 1972. *Karst landforms*. Macmillan, London.

Swinton, W. E. 1970. *The dinosaurs*. George Allen & Unwin, London, 331 pp.

Talbot, M.R. 1971. Calcite cements in the Corallian Beds (Upper Oxfordian) of Southern England. *Journal of Sedimentary Petrology*, 41:261-273.

Thomas, J.B. 1975. The geology of the southern North Sea. *In: Offshore Europe '75*. Offshore Services Magazine, Paper OE-75 213.

Thomas, R.D.K. 1986. Taphonomy: ecology's loss is sedimentology's gain. *Palaios* 1: 206.

Thulborn, R.A. 1973. Teeth of ornithischian dinosaurs from the Upper Jurassic of Portugal. *Memoirs of the Geological Survey of Portugal*, 22: 89-134.

--------- 1991. Morphology, preservation and palaeobiological significance of dinosaur coprolites. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 83: 341-366.

Toots, H. 1965. Random orientation of fossils and its significance. *Contributions to Geology, University of Wyoming*, 4: 59-62.

Torrens, H.S. 1968. The Great Oolite Series. In (Sylvester-Bradley, P.C. & Ford, T.D.; eds.) *Geology of the East Midlands*. Leicester University Press, 227-263.

--------- (ed.) 1969a. *International field symposium on the British Jurassic, Part V (North Somerset, Gloucestershire & South Wales)*. Keele University Press, 35pp.

--------- 1969b. The stratigraphical distribution of Bathonian ammonites in central England. *Geological Magazine*, 106: 63-76.

--------- 1969c. *International field symposium on the British Jurassic, Part I (Dorset and South Somerset)*. Keele University Press, 35pp.

--------- 1974. Standard Zones of the Bathonian. *Colloque du Jurassique a Luxembourg 1967. Memoires du Bureau de Recherches Geologiques et Minieres*. 75:581-604.

Tucker, M.E. *Sedimentary petrology, an introduction*. Geoscience Texts, Volume 3, Blackwell Scientific Publications.

--------- 1990. Skeletal carbonates. *In* (Briggs, D.E.G. & Crowther, P.R.; eds.) *Palaeobiology: a synthesis*. Blackwell, Oxford, 247-250.

--------- & Wright, V.P. 1990. *Carbonate Sedimentology*. Blackwell Scientific Publications.

Trudgill, S. 1985. *Limestone geomorphology. Geomorphology Texts; 8*. Longman, Harlow.

Vail, P.R., Hardenbol, J., & Todd, R.G. 1984. Jurassic unconformities, chronostratigraphy and biostratigraphy. *In* (Schlee, J.S.; ed.) *Interregional Unconformities and Hydrocarbon Accumulation : American Association of Petroleum Geologists Memoir* 36:129-144.

-------- , Mitchum, R.M., JR., Todd, R.G., Widimier, J.M., Thompson, S., III, Sangree, J.B., Blubb, J.N., & Hatlelid, W.G. 1977. Seismic stratigraphy and global sea level, part 2: Application of seismic reflection configuration to stratigraphic interpretations. *In* (Payton, C.E.; ed.) *Seismic Stratigraphy - Applications to Hydrocarbon Exploration. American Association of Petroleum Geologists Memoir* 26:49-212.

--------- , & Todd, R.G. 1981. Northern North Sea Jurassic unconformities, chronostratigraphy and sea-level changes from seismic stratigraphy. *In* (Illing, L.V. & Hobson, G.; eds.) *Petroleum geology of the Continental Shelf of North-West Europe*. Heyden, London. 216-235.

Van Hinte, J.E. 1978. A Jurassic Time Scale. *The American Association of Petroleum Geologists Bulletin*. 60:489-497.

Vaughan, R.F. 1988. Cotswold Dinosaur Excavation. *Geology Today* , 4: 150-151.

---------- 1989. *The excavation at Hornsleasow Quarry. Nature Conservancy Council Interim Report*, 1. Gloucester City Museum. 65pp.

Voorhies, M.R. 1969. Taphonomy and population dynamics of an early Pliocene vertebrate fauna, Knox County, Nebraska. *Contributions in Geology, University of Wyoming, Special Paper 1*. University of Wyoming, Laramie.

--------- 1970. Sampling difficulties in reconstructing late Tertiary mammalian communities. *In* (Yochelson, E.L.; ed.) *Proceedings of the North American palaeontological convention*. Allen Press, Lawrence, 454-468.

Waldman, M. 1974. Megalosaurids from the Bajocian (Middle Jurassic) of Dorset. *Palaeontology*, 17: 325-339.

--------- and Savage, R.G.J. 1972. The first Jurassic mammal from Scotland. *Journal of the Geological Society, London*, 128: 119-125.

--------- and Evans, S.E., 1994. Lepidosauromorph reptiles from the Middle Jurassic of Skye. *Zoological Journal of the Linnean Society*, 112:135-150.

Walford, E.A. On the relation of the so-called "Northampton Sand" of North Oxfordshire to the Clypeus Grit. *Quarterly Journal of the Geological Society of London*, 39: 224-245.

--------- 1895. Stonesfield Slate. - Report of the Committee... *Report of the British Association for the Advancement of Science,* 1894 (1895): 304-306.

--------- 1896. Stonesfield Slate. - Second report of the Committee... *Report of the British Association for the Advancement of Science,* 1895 (1896): 414-415.

--------- 1897. Stonesfield Slate. - Third and final report of the Committee... *Report of the British Association for the Advancement of Science,* 1896 (1897): 356.

--------- 1906. *On some New Oolitic Strata in North Oxfordshire.* The Buckingham Advertiser, Buckingham.

Walkden, G.M. 1974. Palaeokarstic surfaces in upper Visean (Carboniferous) limestones of the Derbyshire Block, England. *Journal of Sedimentary Petrology*, 44: 1232-1247.

Walker, A. D. 1964. Triassic reptiles from the Elgin area: *Ornithosuchus* and the origin of carnosaurs. *Philosophical Transactions of the Royal Society of London, Series B,* 248: 53-134.

Walker, T.R. 1967. Formation of red-beds in modern and ancient deserts. *Bulletin of the Geological Society of America*, 78: 353-368.

Ward, D.J. 1981. A simple machine for bulk processing of clays and silts. *Tertiary Research*, 3: 121-124.

--------- 1984. Collecting isolated microvertebrate fossils. *Zoological Journal of the Linnean Society*, 82: 245-259.

Ward, P.D. 1992. *On Methuselah's Trail. Living fossils and the Great Extinctions.* Freeman and Company, New York, 212pp.

Webster, R. 1965. A horizon of pea-grit in gravel soils. *Nature*, 206: 296-297.

Weigelt, J. 1929. *Recent vertebrate carcasses and their palaeobiological implications.* University of Chicago Press, Chicago (English translation by J. Schaefer)

Weishampel, D. B., Dodson, P. and Osmólska, H. (eds). 1990. *The Dinosauria.* University of California Press, Berkeley, 733 pp.

Weiss, C.P. & Wilkinson, B. 1988. Holocene cementation along the central Texas coast. *Journal of Sedimentary Petrology*, 58:468-478.

Wellnhofer, P. 1978. Pterosauria. *Handbuch der Paläoherpetologie*, 19: 1-82.

--------- 1991. *The illustrated encyclopedia of pterosaurs.* Salamander Books Ltd., London, 192 pp.

Whittacker, A. 1985. *Atlas of Onshore Sedimentary Basins in England and Wales: Post Carboniferous Tectonics and Stratigraphy.* Blackie, London.

Whyte, M.A. & Romano, M. 1993. Footprints of a sauropod dinosaur from the middle Jurassic of Yorkshire. *Proceedings of the Geologists' Association.* 104:195-201.

Wilkinson, B.H. & Landing, E. 1978. "Eggshell diagenesis" and primary radial fabric in calcite ooids. *Journal of Sedimentary Petrology*, 48: 1129-1137.

--------- , Smith, A.L. & Lohmann, K.C. 1985. Sparry calcite marine cement in Upper Jurassic limestones of southeastern Wyoming. In (Schneidermann, N. & Harris, P.M.; eds.) *Carbonate Cements. Special Publications of the Social and Economic Palaeontologists and Mineralogists*, 36: 164-184.

Williams, B.J. & Whittacker, A. 1974. *Geology of the country around Stratford-upon-Avon and Evesham*. Memoir of the British Geological Survey, 151. HMSO, London.

Williams, C.T. 1972. The origin of "spiral coprolites". *University of Kansas Palaeontological Contributions*, 59: 1-19.

--------- 1987. Alteration of chemical composition of fossil bones by soil processes and groundwater. In (Grupe, G. *et al.*; eds.) Trace elements in environmental history. Springer-Verlag, Heidelberg, 27-40.

Wills, M.A. 1993. Annelida. *In* (Benton, M.J.; ed.) *The Fossil Record 2*. Chapman & Hall, London, 271-279.

Wolff, R.G. 1973. Hydrodynamic sorting and ecology of a Pleistocene mammalian assemblage from California (U.S.A.). *Palaeogeography, Palaeoclimatology, Palaeoecology*, 13: 91-101.

Wood, W.R. & Johnson, D.L. 1978. A survey of disturbance processes in archaeological site formation. *In* (Schiffer, M.B.; ed.) *Advances in Archaeological Method and Theory, Volume 1*. Academic Press, New York, 315-381.

Woodhall, D. & Knox, R.W.O'B. 1979. Mesozoic volcanism in the northern North Sea and adjacent areas. *Bulletin of the Geological Survey of Great Britain*, 70: 34-76.

Woodward, A. S. 1890. A synopsis of the fossil fishes of the English Lower Oolites. *Proceedings of the Geologists' Association*, 11: 285-306.

--------- 1916-1919. *The fossil fishes of the English Wealden and Purbeck formations*. Palaeontographical Society Monograph, London.

Woodward, H. B. 1894. *The Jurassic rocks of Britain. Vol. IV. The Lower Oolitic rocks of England (Yorkshire excepted)*. HMSO, London.

--------- 1895. *The Jurassic rocks of Britain. Vol. V. The Middle and Upper Oolitic rocks of England (Yorkshire excepted)*. HMSO, London.

Worthy, T.H. & Holdaway, R.N. 1994. Scrap's form an owl's table - predator activity as a significant taphonomic process newly recognised from New Zealand Quaternary deposits. *Alcheringa*, 18: 229-245.

Wright, D.K. 1872. On the correlation of the Jurassic rocks in the department of the Cote-d'Or, France, with the Oolitic Formations in the counties of Gloucester and Wiltshire. *Proceedings of the Cotteswold Naturalist Field Club*. 5:143-238.

Wright, V.P. 1982. The recognition and interpretation of palaeokarsts: two examples from the Lower Carboniferous of South Wales. *Journal of Sedimentary Petrology*. 5: 83-94.

--------- 1983. A rendzina from the Lower Carboniferous of South Wales. *Sedimentology*, 30: 159-179.

--------- 1986. *Paleosols: their recognition and interpretation*. Oxford, Blackwell.

--------- 1987. The ecology of two early Carboniferous paleosols. *In* (Miller, J., Adams, A.E. & Wright, V.P.; eds.) *European Dinantian Environments*. Geological Journal Special Issue, 12: 345-358.

--------- & Peeters, C. 1989. Origins of some early Carboniferous calcrete fabrics revealed by cathodeluminescence: implications for interpreting the sites of calcrete formation. *Sedimentary Geology*, 65: 345-353.

--------- & Wilson, R.C.L. 1987. A terra rossa-like paleosol complex from the Upper Jurassic of Portugal. *Sedimentology*, 34: 259-274.

Wyatt, R.J. 1981. *Geological notes and local details for 1:10 000 sheet SP31NE (Charlbury and Stonesfield)*. Institute of Geological Sciences, Keyworth.

Yaalon, D.H. 1971. Soil-forming processes in time and space. *In* (Yaalon, D.H.; ed.) *Paleopedology*. Israel University Press, Jerusalem.

Young, T. 1984. Neoselachian sharks from the Middle Jurassic of England. *Journal of the University of Sheffield Geological Society*, 8: 24-31.

Ziegler, P.A. 1981. Evolution of sedimentary basins in North-West Europe. *In* (Illing, L.V. & Hobson, G.; eds.) *Petroleum geology of the Continental Shelf of North-West Europe*. Heyden, London. 3-39.

Appendix A: British Fossil vertebrate sites

Locality	Area	U.K. N.G.R	Group	Formation	Member	Zone	Subzone	Environment	Fish	Rept.	Mam.	Foot.	Status	Notes	Reference
Eype	Dorset	SY4592	Inferior Oolite					marine		x			cliff-section	teleosaur	Benton & Spencer 1995
Crewbourne	Dorset	ST458104	Inferior Oolite	?Astarte Bed		*?garantiana*	*?acris*	marine		x			unknown	ichthyosaur	Benton & Spencer 1995
Bradford Abbas	Dorset	SY5915	Inferior Oolite					marine		x			infilled	stegosaur	Benton & Spencer 1995
Nethercombe Qu.	Dorset	ST636175	Inferior Oolite	Red conglomerate		*humphriesianum*		proximal marine		x			infilled	TYPE (megalosaur)	Benton & Spencer 1995
Cold Harbour Road Qu	Dorset	ST6417	Inferior Oolite	Burton limestone		*parkinsoni*		marine		x			infilled	TYPE (megalosaur)	Benton & Spencer 1995
Doulting Qus.	Somer.	ST6543	Inferior Oolite	?Doulting Stone		*?parkinsoni*		?marine		x			unknown	megalosaur	Benton & Spencer 1995
Stroud railway cutting	Glos.	SO8505	Inferior Oolite					?marine		x			cutting	megalosaur	Benton & Spencer 1995
Frith Qu.	Glos.	SO868083	Inferior Oolite	Lower Trigonia Grit		*discites*		marine		x			infilled	plesiosaur	Benton & Spencer 1995
Rodborough Hill	Glos.	SO8404	Inferior Oolite					?marine		x			cutting	teleosaur	Benton & Spencer 1995
Leckhampton Qus.	Glos.	SO950185	Inferior Oolite	Gryphite Grit		*laeviuscula*	*?ovalis*	marine	x	x			disused	marine reptiles	Benton & Spencer 1995
Crickley Hill Qu.	Glos.	SO928164	Inferior Oolite					?marine		x			disused	megalosaur	Benton & Spencer 1995
Cornwell	Oxon.	SP267271	Inferior Oolite	?Clypeus Grit		*parkinsoni*		marine		x			infilled	megalosaur	Plot 1677
Saltwick Bay	Yorks.	NZ 9110	Ravenscar	Hayburn		*murchisonae*	*?murchisonae*	fluvio-deltaic				x	cliff-section	trackways	Whyte & Romano1993
nr. Whitby shore	Yorks.	NZ909113	Ravenscar	Hayburn		*murchisonae*	*?murchisonae*	fluvio-deltaic				x	cliff-section	sauropod track	Whyte & Romano1993
Peak Alum Qus.	Yorks.	NT973017	Ravenscar	Hayburn		*murchisonae*	*?murchisonae*	fluvio-deltaic				x	disused	footprint	Delair & Sargeant 1985
Cloughton Wyke	Yorks.	TA020950	Ravenscar	Cloughton	Gristhorpe	*sauzei*		alluvial				x	fallen blocks	footprints	Delair & Sargeant 1985
White Nab shore	Yorks.	TA058864	Ravenscar	Scarborough		*humphriesianum*	*?blagdeni*	marine	x	x			beach section	trackways	Delair & Sargeant 1985
Cromer Point	Yorks.	TA031928	Ravenscar	Scalby	Long Nab	*parkinsoni*	*?bomfordi*	alluvial				x	cliff-section	trackways	Delair & Sarjeant 1985
Hundale Point	Yorks.	TA027949	Ravenscar	Scalby	Moor Grit	*?subfurcatum-parkinsoni*		alluvial				x	beach section	footprints	Delair & Sargeant 1985

Table A1. British Aalenian and Bajocian fossil vertebrate sites

Locality	Area	U.K. N.G.R	Group	Formation	Member	Zone	Subzone	Environment	Fish	Amph.	Rept.	Mam.	Foot.	Status	Notes	Reference
Smith's Qu.	Oxon.	SP300226	Great Oolite	Chipping Norton	Chipping Norton	*zigzag*	*macrescens*	?karst	x		x			disused	microvertebrates	Evans & Milner 1994
Padley's Qu.	Oxon.	SP317269	Great Oolite	Chipping Norton		*zigzag*		proximal marine			x			overgrown	cetiosaur	Woodward 1894
Workhouse Qu.	Oxon.	SP319274	Great Oolite	Sharps Hill		*tenuiplicatus*		lagoonal			x			disused	reptile fauna	Benton & Spencer 1995
Chapel House	Oxon.	SP329282	Great Oolite	Chipping Norton		*zigzag*		?marine			x			infilled	cetiosaur	Phillips 1871
Cross Hands Qu.	Oxon.	SP269292	Great Oolite	Chipping Norton		*zigzag*		marine	x					landfill	microvertebrates	pers obs.
nr. Burford	Oxon.	SP2511	Great Oolite	?Charlbury	Eyford	*progracilis*		subaerial					x	unknown	mammal footprint	Delair & Sargeant 1985
Over Norton Qu.	Oxon.	SP315290	Great Oolite	Chipping Norton		*zigzag*		proximal marine			x			infilled	cetiosaur	Benton & Spencer 1995
Enstone Qu.	Oxon.	SP383250	Great Oolite	Chipping Norton		*zigzag*		proximal marine			x			landfill	cetiosaur	Phillips 1871
Temple Mills Qu.	Oxon.	SP347363	Great Oolite	Sharps Hill		*tenuiplicatus*		lagoonal			x			overgrown	steneosaur	Benton & Spencer 1995
Sharps Hill Qu.	Oxon.	SP337358	Great Oolite	Sharps Hill		*tenuiplicatus*		lagoonal	x		x			overgrown	diverse fauna	Evans & Milner 1994
Stonesfield Mines	Oxon.	SP387171	Great Oolite	Taynton	'Stonesfield Slates'	*progracilis*		proximal marine	x		x	x		disused	SSSI; diverse fauna	Woodward 1894
Slape Hill Qu.	Oxon.	SP425195	Great Oolite	White Limestone		*subcontractus-hodsoni*		marine			x			active	crocodile fauna	Palmer pers. comm.
Glympton Qu.	Oxon.	SP427217	Great Oolite	Forest Marble		*aspidoides-discus*		proximal marine			x			infilled	cetiosaur	Phillips 1871
Tolley's Qu.	Oxon.	SP4414	Great Oolite					?marine			x			unknown	reptile fauna	Benton & Spencer 1995
Hanborough Station	Oxon.	SP430144	Great Oolite	Lower Combrash		*discus*	*discus*	marine			x			cutting	crocodile	Benton & Spencer 1995
Hampton Common	Oxon.	SP5015	Great Oolite					?marine			x			unknown	reptile fauna	Benton & Spencer 1995
Woodeaton Qu.	Oxon.	SP533124	Great Oolite	Hampen Marly		*progracilis*		lagoonal	x	x	x	?x		disused	diverse fauna	Evans & Milner 1994
Enslow Bridge Qu.	Oxon.	SP478187	Great Oolite	Taynton	'Stonesfield Slates'	*progracilis*		proximal marine			x			disused	TYPE steneosaur	Benton & Spencer 1995
Enslow Bridge Qu.	Oxon.	SP478187	Great Oolite	Forest Marble		*aspidoides-discus*		lagoonal	x		x			disused	dinosaur fauna	Benton & Spencer 1995
Bletchingdon Station Qu.	Oxon.	SP483181	Great Oolite	White Limestone	Bladon	*hodsoni*		coastal marsh	x		x			cutting	reptile fauna	Benton & Spencer 1995
Shipton Cement Works	Oxon.	SP475175	Great Oolite	White Limestone	Bladon	*hodsoni*		coastal marsh	x		x			active	SSSI; diverse fauna	Evans & Milner 1994
Shipton Cement Works	Oxon.	SP475175	Great Oolite	Forest Marble		*aspidoides-discus*		lagoonal	x		x			active	SSSI; diverse fauna	Evans & Milner 1994
Kirtlington Cement Works	Oxon.	SP494199	Great Oolite	White Limestone	Bladon	*hodsoni*		coastal marsh			x			disused	SSSI; diverse fauna	Evans & Milner 1994
Kirtlington Cement Works	Oxon.	SP494199	Great Oolite	Forest Marble	'Mammal bed'	*aspidoides-discus*		lagoonal	x	x	x	x		disused	SSSI; microvertes.	Evans & Milner 1994
nr. Oxford	Oxon.	SP5106	Great Oolite					?marine			x			unknown	steneosaur	Benton & Spencer 1995
Littlemore railway	Oxon.	SP5202	Great Oolite	Lower Combrash		*discus*	*discus*	marine			x			cutting	megalosaur	Benton & Spencer 1995
Fritwell railway	Oxon.	SP5228	Great Oolite					?marine			x			cutting	teleosaur	Benton & Spencer 1995
Ardley Field Farm Qu.	Oxon.	SP544264	Great Oolite	White Limestone	Ardley	*hodsoni*		proximal marine	x		x			active	reptile fauna	Benton & Spencer 1995
Ardley Cutting Qu.	Oxon.	SP538273	Great Oolite	White Limestone	Ardley	*hodsoni*		proximal marine			x			overgrown	crocodile fauna	Benton & Spencer 1995
Ardley Cutting Qu.	Oxon.	SP538273	Great Oolite	White Limestone	Shipton	*morrisi*		subaerial				x		overgrown	trackway	Benton & Spencer 1995
Stratton Audley Qus.	Oxon.	SP603252	Great Oolite	Forest Marble		*aspidoides-discus*		lagoonal			x			lakes	reptile fauna	Benton & Spencer 1995
Blisworth railway	North.	SP725543	Great Oolite	Blisworth Limestone		*hodsoni*		proximal marine			x			cutting	reptile fauna	Benton & Spencer 1995
Cogenhoe	North.	SP8360	Great Oolite	Lower Combrash		*discus*	*discus*	marine			x			unknown	cetiosaur	Benton & Spencer 1995
Buttock's Booth	North.	SP7864	Great Oolite	Blisworth Limestone		*hodsoni*		proximal marine			x			infilled	steneosaur	Benton & Spencer 1995
Kingsthorpe	North.	SP7563	Great Oolite					?marine			x			unknown	steneosaur	Benton & Spencer 1995
Rushden Qu.	North.	SP951661	Great Oolite	Lower Combrash		*discus*	*discus*	marine			x			disused	plesiosaur	Benton & Spencer 1995
Irchester	North.	SP8968	Great Oolite					?marine			x			unknown	steneosaur	Benton & Spencer 1995

Table A2. British Bathonian fossil vertebrate sites

Locality	Area	U.K. N.G.R	Group	Formation	Member	Zone	Subzone	Environment	Fish	Amph.	Rept.	Mam.	Foot.	Status	Notes	Reference
Long Burton	Dorset	ST6513	Great Oolite	?Forest Marble		*aspidoides-discus*		proximal marine			x			infilled	megalosaur	Benton & Spencer 1995
Yetminster	Dorset		Great Oolite	Lower Combrash	Astarte Bed	*discus*	*discus*	marine			x			unknown	ophthalmosaur	Benton & Spencer 1995
Watton Cliff	Dorset	SY451908	Great Oolite	Forest Marble		*aspidoides-discus*		proximal marine	x	x	x	x		cliff section	SSSI; microvertes	Kermack 1988
Swyre	Dorset	SY525868	Great Oolite	Forest Marble		*aspidoides-discus*		proximal marine	x	x	x	x		exposed	microvertebrates	Evans & Milner 1994
Shipmoor	Dorset	SY576836	Great Oolite	Forest Marble		*aspidoides-discus*		proximal marine	x					beach section	microvertebrates	pers. obs.
Closworth	Somer.	ST5610	Great Oolite	Lower Combrash		*discus*	*discus*	marine			x			unknown	TYPE-steneosaur	Benton & Spencer 1995
Bath	Avon	ST7565	Great Oolite	Forest Marble	'Great Oolite'	*aspidoides*		marine			x			unknown	TYPE-steneosaur	Benton & Spencer 1995
Bradford-Avon	Wilts.	ST8260	Great Oolite	Forest Marble		*discus*	*hollandi*	proximal marine			x			unknown	cetiosaur	Benton & Spencer 1995
Atford	Wilts.	ST8666	Great Oolite	Forest Marble		*aspidoides-discus*		proximal marine	x		x			infilled	reptile fauna	Benton & Spencer 1995
Malmesbury	Wilts.	ST9387	Great Oolite								x			unknown	cetiosaur	Benton & Spencer 1995
Avoncliff	Wilts.	ST8059	Great Oolite	Lower Fullers Earth		*zigzag-progracilis*		marine			x			disused	teleosaur	Benton & Spencer 1995
Frankley	Wilts.		Great Oolite								x			unknown	cetiosaur	Benton & Spencer 1995
Box Tunnel	Wilts	ST8469	Great Oolite								x			unknown	megalosaur	Benton & Spencer 1995
Leigh Delamere	Wilts.	ST890790	Great Oolite	Forest Marble		*aspidoides-discus*		proximal marine	x	x	x	x		exposed	microvertebrates	Evans & Milner 1994
Sutton Benger	Wilts.		Great Oolite	Forest Marble		*aspidoides-discus*		?lagoonal					x	unknown	quadroped	Sarjeant 1974
Minchinhampton res.	Glos.	SO871003	Great Oolite	White Limestone		*subcontractus-hodsoni*		marine			x			reservoir	TYPE proceratosaur	Weishampel et al. 1990
Avening	Glos.	ST8897	Great Oolite	Forest Marble		*aspidoides-discus*		marine			x			unknown	teleosaur	Benton & Spencer 1995
Tarlton Clay Pit	Glos.	SO970001	Great Oolite	Forest Marble		*aspidoides-discus*		marine	x	x	x	x		active	microvertebrates	Evans & Milner 1994
Sapperton Tunnel	Glos.	SO951015	Great Oolite	White Limestone	Kemble Beds	*hodsoni*		marine			x			railway	reptile fauna	Benton & Spencer 1995
Jarvis Old Qu.	Glos.	ST995999	Great Oolite	White Limestone	Kemble Beds	*hodsoni*		marine	x		x			rifle range	reptile fauna	Benton & Spencer 1995
Shorncote Hill Qu.	Glos.	SU032972	Great Oolite	Forest Marble		*aspidoides-discus*		proximal marine	x		x			exposed	microvertebrates	pers. obs
Hare Bushes Qu.	Glos.	SP0303	Great Oolite	White Limestone	?Kemble Beds	*hodsoni*		marine			x			infilled	?teleosaur 'eggs'	Woodward 1894
Ampney Downs Qu.	Glos.	SP047068	Great Oolite	White Limestone		*subcontractus-hodsoni*		marine	x					active	sharks teeth	pers. obs
Daglingworth Qu.	Glos.	SP000060	Great Oolite	White Limestone	Shipton	*subcontractus-morrisi*		proximal marine			x			active	reptile fauna	pers obs
Ready Token	Glos.	SP100050	Great Oolite	Forest Marble		*aspidoides-discus*		proximal marine	x		x	x		?active	microvertebrates	Evans & Milner 1994
Bibury	Glos.	SP0611	Great Oolite	Forest Marble		*aspidoides-discus*		proximal marine			x			infilled	cetiosaur	Benton & Spencer 1995
Chedworth cutting	Glos.	SP052135	Great Oolite	White Limestone	Shipton	*subcontractus-morrisi*		marine			x			overgrown	steneosaur	Woodward 1894
Sevenhampton Qus.	Glos.	SP0122	Great Oolite	Charlbury	Eyford	*progracilis*		proximal marine	x		x			infilled	diverse fauna	pers. obs
Salperton Qu.	Glos.	SP078211	Great Oolite	Charlbury	Eyford	*progracilis*		proximal marine	x					disused	microvertebrates	pers. obs
Brockhill Qu.	Glos.	SP133236	Great Oolite	Charlbury	Eyford	*progracilis*		proximal marine	x		x			active	diverse fauna	pers. obs.
Huntsmans Qu.	Glos.	SP1225	Great Oolite	Charlbury	Eyford	*progracilis*		proximal marine	x		x			active	SSSI; diverse fauna	pers. obs.
Kineton Thorns Qu.	Glos.	SP1226	Great Oolite	Charlbury	Eyford	*progracilis*		proximal marine	x		x			infilled	diverse fauna	pers. obs.
Eyford Hill Qu.	Glos.	SP131252	Great Oolite	Charlbury	Eyford	*progracilis*		proximal marine	x		x			infilled	Witts collection	pers. obs.
New Park Qu.	Glos.	SP176292	Great Oolite	Chipping Norton	Hook Norton	*zigzag*	*convergens*	lagoonal	x		x			disused	SSSI; diverse fauna	Reynolds 1939
Longborough Road Qu.	Glos.	SP171295	Great Oolite	Chipping Norton	?Hook Norton	*zigzag*	*?convergens*	?lagoonal			x			infilled	reptile fauna	Arkell 1933
Hornsleasow Qu.	Glos.	SP131322	Great Oolite	Chipping Norton	Chipping Norton	*zigzag*	*macrescens*	karst	x	x	x	x		active	SSSI; diverse fauna	pers. obs.
Oakham Qu.	Glos.	SP279306	Great Oolite	Chipping Norton		*zigzag*		?terrestrial			x			disused	reptiles fauna	Benton & Spencer 1995
Stanton cutting	Glos.	SP0634	Great Oolite	Forest Marble		*aspidoides-discus*		proximal marine			x			overgrown	megalosaur	Benton & Spencer 1995

Table A2, continued....

Locality	Area	U.K. N.G.R	Group	Formation	Member	Zone	Subzone	Environment	Fish	Amph.	Rept.	Mam.	Foot.	Status	Notes	Reference
Islip Ironstone Qu.	North	SP975782	Great Oolite	Lower Combrash		*discus*	*discus*	marine			x			disused	reptile fauna	Benton & Spencer 1995
Oundle	North.	TL0388	Great Oolite					?marine			x			unknown	crocodile	Benton & Spencer 1995
Olney	Bucks.	SP8851	Great Oolite	Lower Combrash		*discus*	*discus*	marine			x			unknown	reptile fauna	Benton & Spencer 1995
Thomborough old mill	Bucks.	SP	Great Oolite	Forest Marble		*aspidoides-discus*		subaerial					x	field	reptile footprint	Delair & Sargeant 1985
nr. Peterborough	Cambs.	TL1998	Great Oolite	Lower Combrash		*discus*	*discus*	marine			x			unknown	crocodiles	Benton & Spencer 1995
Orton	Cambs.	TL1795	Great Oolite					?marine			x			unknown	crocodile	Benton & Spencer 1995
Norman Cross Brickworks	Cambs.	TL170912	Great Oolite	Lower Combrash		*discus*	*discus*	marine			x			active	reptile fauna	Benton & Spencer 1995
Banthorpe railway	Leics.	TL0412	Great Oolite	Blisworth Clay		*aspidoides-discus*		lagoonal			x			cutting	reptile fauna	Benton & Spencer 1995
Belmesthorpe	Leics.	TL0410	Great Oolite	Rutland		*zigzag-progracilis*		coastal marshes	x		x			unknown	diverse fauna	Benton & Spencer 1995
Black Rocks shore	Yorks.	TA055865	Ravenscar	Scalby	Long Nab	no correlation		alluvial					x	cliff section	footprints	Benton & Spencer 1995
White Nab shore	Yorks.	TA060862	Ravenscar	Scalby	Long Nab	no correlation		alluvial					x	exposed	trackways	pers. obs.
Burniston Wyke	Yorks.	TA 028935	Ravenscar	Scalby	Long Nab	no correlation		alluvial					x	cliff section	'Footprint Bed'	Sargeant 1974
Long Nab shore	Yorks.	TA031940	Ravenscar	Scalby	Long Nab	no correlation		alluvial					x	cliff section	footprints	pers. obs.
Kildonnan shore	Eigg	NM495870	Great Estuarine	Lealt Shale	Kildonnan	*zigzag*	*?macrescens*	lagoonal	x		x			beach section	SSSI; diverse fauna	Hudson 1966
Eilean Thuilm	Eigg	NM483913	Great Estuarine	Lealt Shale	Kildonnan	*zigzag*	*?macrescens*	lagoonal	x		x			beach section	SSSI; diverse fauna	Hudson 1966
Port na Cullaidh shore	Skye	NG516136	Great Estuarine	Cullaidh Shale		*zigzag*	*convergens*	marine	x					beach section	articulated fish	Hudson 1962
Bidein an Fhithich cliffs	Skye	NG517147	Great Estuarine	Kilmaluag		*?hodsoni-aspidoides*		lagoonal	x		x			cliff section	SSSI; microvertes.	Andrews 1985
South Cam Mor shore	Skye	NG518154	Great Estuarine	Kilmaluag		*?hodsoni-aspidoides*		lagoonal	x		x			beach section	SSSI; microvertes.	Andrews 1985
Cladach a' Ghlinne	Skye	NG520161	Great Estuarine	Kilmaluag		*?hodsoni-aspidoides*		lagoonal	x		x	x		beach section	SSSI; microvertes.	Waldman & Savage 1972
Cladach a' Ghlinne	Skye	NG520161	Great Estuarine	Duntulm		*?morrisi*		lagoonal	x					beach section	fish teeth	Andrews & Walton 1990
North Glen Scaladal shore	Skye	NG518168	Great Estuarine	Kilmaluag		*?hodsoni-aspidoides*		lagoonal	x		x	x		exposures	SSSI; microvertes.	Waldman & Savage 1972
Bearreraig Bay	Skye	NG518525	Great Estuarine	Cullaidh Shale		*zigzag*	*convergens*	marine	x					cliff section	articulated fish	Hudson 1962
Rubha nam Brathairean	Skye	NG526625	Great Estuarine	Lealt Shale	Lonfearn	*?zigzag-tenuiplicatus*		subaerial	x				x	cliff section	dinosaur footprint	Andrews & Hudson 1984
Valtos beach	Skye	NG516638	Great Estuarine	Valtos Sandstone		*tenuiplicatus-morrisi*		lagoonal-deltaic	x					beach section	fish fauna	Harris & Hudson 1980
Dun Dearg Qu.	Skye	NG514644	Great Estuarine	Valtos Sandstone		*tenuiplicatus-morrisi*		lagoonal-deltaic	x					overgrown	fish fauna	Harris & Hudson 1980
Digg shore	Skye	NG474694	Great Estuarine	Skudiburgh		*aspidoides*		alluvial	x					cliff section	microvertebrates	Andrews 1985
Kilmaluag Bay	Skye	NG4475	Great Estuarine	Kilmaluag		*?hodsoni-aspidoides*		lagoonal	x		?x			beach section	4 fish horizons	Andrews 1985
Cairdh Ghlumaig bay	Skye	NG410740	Great Estuarine	Duntulm		*?morrisi*		lagoonal	x		x			beach section	microvertebrates	Andrews & Walton 1990

Table A2. continued....

Locality	Area	U.K. N.G.R	Group	Formation	Member	Zone	Subzone	Environment	Fish	Amph.	Rept.	Mam.	Foot.	Status	Notes	Reference
Weymouth backwater	Dorset	SY677790	Oxford Clay	Oxford Clay	Lower	*calloviense-athleta*		marine	x		x			beach section	reptile fauna	Benton & Spencer 1995
Chickerell Brickyard	Dorset	SY644797	Oxford Clay	Oxford Clay	Lower	*calloviense-athleta*		marine	x		x			disused	reptile fauna	Martill & Hudson 1991
Putton Lane Brick Pit	Dorset	SY650801	Oxford Clay	Oxford Clay	Lower	*?jason*		marine	x		x			disused	TYPE plesiosaur	Benton & Spencer 1995
Bowleaze Cove	Dorset	SY702818	Oxford Clay	Oxford Clay	Middle	*?lamberti*	*?lamberti*	marine			x			cliff section	reptile fauna	Benton & Spencer 1995
Radipole	Dorset	SY6781	Oxford Clay	Oxford Clay	?Lower	*?calloviense-athleta*		marine			x			unknown	plesiosaur	Benton & Spencer 1995
Tidmoor Point, Fleet	Dorset	SY643786	Oxford Clay	Oxford Clay	Middle	*athleta-lamberti*		marine	x		x			beach section	diverse fauna	Martill & Hudson 1991
Devizes	Wilts.	SU0661	Oxford Clay	Oxford Clay		*calloviense-lamberti*		marine			x			unknown	plesiosaur	Benton & Spencer 1995
Melksham	Wilts.	ST9063	Oxford Clay	Oxford Clay		*calloviense-lamberti*		marine			x			unknown	reptile fauna	Benton & Spencer 1995
Chippenham	Wilts.	ST9173	Oxford Clay	Oxford Clay		*calloviense-lamberti*		marine	x		x			unknown	diverse fauna	Benton & Spencer 1995
Christian Malford tnal pits	Wilts.	ST956775	Oxford Clay	Oxford Clay	Lower	*athleta*	*phaeinum*	marine	x		x			infilled	lagerstatten	Martill & Hudson 1991
Long Marston	Oxon.	SP5309	Oxford Clay	Oxford Clay		*calloviense-lamberti*		marine			x			unknown	reptile fauna	Benton & Spencer 1995
Shotover Hill	Oxon.	SP5706	Oxford Clay	Oxford Clay		*calloviense-lamberti*		marine			x			unknown	reptile fauna	Benton & Spencer 1995
Cowley Field Pit	Oxon.	SP5703	Oxford Clay	Oxford Clay		*calloviense-lamberti*		marine	x		x			infilled	TYPE plesiosaur	Benton & Spencer 1995
St. Clements pits	Oxon.	SP5306	Oxford Clay	Oxford Clay		*calloviense-lamberti*		marine			x			infilled	reptile fauna	Benton & Spencer 1995
Calvert Brick Works	Oxon.	SP695234	Oxford Clay	Oxford Clay	Lower-Middle	*calloviense-athleta*		marine	x					active	fish fauna	Martill & Hudson 1991
Summertown Brick Pit	Oxon.	SP510090	Oxford Clay	Oxford Clay	Middle	*athleta*		marine	x		x			infilled	diverse fauna	Martill & Hudson 1991
Wolvercote Brickpit	Oxon.	SP494105	Oxford Clay	Oxford Clay	Lower-Middle	*calloviense-athleta*		marine	x		x			infilled	diverse fauna	Benton & Spencer 1995
Shellingford Crossroads Qu.	Oxon.	SU326942	Oxford Clay	Oxford Clay		*calloviense-lamberti*		marine			x			infilled	reptile fauna	Benton & Spencer 1995
St. Edmund Hall	Oxon.	SP518063	Oxford Clay	Oxford Clay	Middle	*lamberti*	*henrici*	marine			x			infilled	crocodile	Benton & Spencer 1995
Iffley Road sports ground	Oxon.	SP523053	Oxford Clay	Oxford Clay		*calloviense-lamberti*		marine			x			infilled	plesiosaur	Benton & Spencer 1995
Cumnor	Oxon.	SP465065	Oxford Clay	Oxford Clay		*calloviense-lamberti*		marine			x			?cutting	plesiosaur	Benton & Spencer 1995
Buckingham	Bucks.	SP6933	Oxford Clay	Oxford Clay		*calloviense-lamberti*		marine			x			unknown	icthyosaur	Benton & Spencer 1995
Bletchley Brick Works	Bucks.	SP855315	Oxford Clay	Oxford Clay	Lower-Middle	*jason-athleta*		marine			x			active	icthyosaur	Martill & Hudson 1991
Caldecotte reservoir	Bucks.	SP892352	Oxford Clay	Oxford Clay	Lower	*coronatum*		marine	x		x			lake	diverse fauna	Martill & Hudson 1991
Newton Longville pits	Bucks.	SP853322	Oxford Clay	Oxford Clay	Lower	*calloviense-athleta*		marine			x			active	pliosaur	Benton & Spencer 1995
Penny Stratford	Bucks.	SP8834	Oxford Clay	Oxford Clay		*calloviense-lamberti*		marine			x			unknown	steneosaur	Benton & Spencer 1995
Martson Moretaine pits	Beds.	TL005425	Oxford Clay	Oxford Clay	Lower	*calloviense-athleta*		marine			x			disused	pliosaur	Martill & Hudson 1991
Stewartby Clay Pit	Beds.	TL030420	Oxford Clay	Oxford Clay	Lower	*calloviense*	*enodatum*	marine	x		x			active	diverse fauna	Martill & Hudson 1991
Kempston clay pits	Beds.	TL036450	Oxford Clay	Oxford Clay	Lower	*calloviense-athleta*		marine	x		x			disused	diverse fauna	Martill & Hudson 1991
Bedford	Beds.	TL0449	Oxford Clay	Oxford Clay	Lower	*calloviense-athleta*		marine	x		x			unknown	diverse fauna	Benton & Spencer 1995
Ravensden	Beds.	TL0754	Oxford Clay	Oxford Clay	Lower	*calloviense-athleta*		marine			x			unknown	pliosaur	Benton & Spencer 1995
Eynesbury Brick Works	Cambs.	TL1859	Oxford Clay	Oxford Clay	Lower	*calloviense-athleta*		marine			x			disused	TYPE pliosaur	Benton & Spencer 1995
St. Neots Brickyard	Cambs.	TL1860	Oxford Clay	Oxford Clay	Lower	*coronatum*		marine			x			disused	pliosaur	Benton & Spencer 1995
St. Ives Brickyard	Cambs.	TL326715	Oxford Clay	Oxford Clay	Middle	*athleta-lamberti*		marine	x		x			disused	diverse fauna	Martill & Hudson 1991
Buntings Lane Borrow Pit	Cambs.	TL200957	Oxford Clay	Oxford Clay	Lower	*calloviense*	*enodatum*	marine	x		x			dump	diverse fauna	Martill & Hudson 1991
Buntings Lane Borrow Pit	Cambs.	TL200957	Kellaways	Kellaways Rock		*calloviense*	*calloviense*	shallow marine	x					dump	fish fauna	Martill & Hudson 1991

Table A3. British Callovian fossil vertebrate sites

Locality	Area	U.K. N.G.R	Group	Formation	Member	Zone	Subzone	Environment	Fish	Amph.	Rept.	Mam.	Foot.	Status	Notes	Reference
Dogsthorpe Brick Pit	Cambs.	TL210020	Oxford Clay	Oxford Clay	Lower	*jason*		marine	x		x			disused	diverse fauna	Martill & Hudson 1991
Orton Brick Pit	Cambs.	TL170940	Oxford Clay	Oxford Clay	Lower	*jason-coronatum*		marine	x					active	fish fauna	Martill & Hudson 1991
Woodston Lodge	Cambs.	TL1897	Oxford Clay	Oxford Clay	Lower	*calloviense-athleta*		marine			x			unknown	reptile fauna	Benton & Spencer 1995
Fletton Brick Works	Cambs.	TL1995	Oxford Clay	Oxford Clay	Lower	*jason*		marine	x		x			active	diverse fauna	Benton & Spencer 1995
Peterborough Gas Works	Cambs.	TL199991	Kellaways	Kellaways Sand		*calloviense*	*koenigi*	marine			x			infilled	TYPE sauropod	Martill 1986
Barrow Pit	Cambs.	TL200958	Oxford Clay	Oxford Clay	Lower	*calloviense-athleta*		marine	x		x			disused	diverse fauna	Benton & Spencer 1995
Stanground	Cambs.	TL2097	Oxford Clay	Oxford Clay	Lower	*jason*		marine			x			infilled	reptile fauna	Benton & Spencer 1995
Eye	Cambs.	TL2202	Oxford Clay	Oxford Clay	Lower	*calloviense-athleta*		marine			x			unknown	steneosaur	Benton & Spencer 1995
Whittlesey brick pit	Cambs.	TL250967	Oxford Clay	Oxford Clay	Lower	*jason*		marine	x		x			active	diverse fauna	Martill & Hudson 1991
Whittlesey brick pit	Cambs.	TL247975	Oxford Clay	Oxford Clay	Lower	*jason*		marine	x		x			active	diverse fauna	Martill & Hudson 1991
Reepham	Lincs.	TF046747	Kellaways	Kellaways Sand		*calloviense*		shallow marine	x		x			trial pit	diverse fauna	Brown & Keen 1991
Mill Hill, Elloughton	Humbs.	SE942278	Kellaways	Kellaways Sand		*calloviense*		shallow marine	x		x			trial pit	diverse fauna	Brown & Keen 1991
Hackness	Yorks.	SE9690	Kellaways	Kellaways Sand		*calloviense*		shallow marine			x			disused	reptile fauna	Benton & Spencer 1995
Gristhorpe	Yorks.	TA1283	Kellaways	Kellaways Rock		*calloviense*		shallow marine			x			unknown	steneosaur	Benton & Spencer 1995

Table A3. continued....

APPENDIX B1: Section at Hornsleasow Quarry as defined by Torrens (1969)

" **5. TAYNTON STONE.** Flaggy, buff oolitic limestone full of comminuted shell debris. Upper surfaces ripple marked. A rather rich fauna includes two *Procerites* sp. indet indicating *Progracilis* Zone. *Procymatoceras* sp., *Trigonia (Myophorella) impressa* J. de C. Sow., *Liostera hebridica* Forbes, *Strophodus* teeth and fragments of wood

seen 2.5-3.0m.

4. SHARPS HILL BEDS

4?j Upper 'Nerinaea' Bed. Brown, unfossiliferous, oolitic clay at base passing up into grey green fine clay with poorly preserved large *Aphanoptyxis* sp.

top not seen over 0.45m.

4?i Very soft, rubbly, oolitic, pink limestone becoming flaggy near top, numerous vertical ?worm burrows

0.3m.

4?h Middle Nerinaea Bed. Grey green, soft, rubbly, oolitic marl passing down into orange brown oolitic clay with *Aphanoptyxis* sp. and *Liostrea* sp.

0.45m.

4?g Shelly, oolitic limestone (Channon's Bed 7 - "Stonesfield Slate"), cross-bedded, hard near base but more flaggy weathering near top. Much shell and *Liostrea* debris but macrofossils few and fragmentary

2.5m.

4f Blue shaley clays grading up into brown friable marl, at the base a different fauna of gastropods from those seen below, including common large "*Natica*" casts, *Angaria* n. sp. (teste L.R. Cox), and an unidentified high spired genus represented by large internal moulds; *Kallirhynchia* sp. and *Modiolus imbricatus* J. Sow. also present

0.5m.

4e Coral Bed A mass of finely preserved corals set in a blue clay matrix. *Microsolena excelsa* Edwards and Haime very abundant, and less commonly, *Cyathopora pratti* Edwards and Haime, *Isastrea limitata* (Lamouroux) and *Thamastrea lyelli* (Edwards and Haime). Many of the corals are overgrown with serpulids and almost all heavily bored by *Lithophaga*, evidence of a marked pause in sedimentation at this horizon. Also present are common crushed *Kallirhynchia* sp. in original life position fixed by a pedicle to the corals, frequent *Modiolus imbricatus* J. Sow. often similarly in original byssally fixed life position and oysters (*Liostrea*) attached to the corals. *Epithyris* sp. also present. Some of the corals found here in

the past have reached 50cms. in diameter and although internally recrystallized, show superbly preserved growth banding and much variation in morphology between corallites thickness variable

0.1-0.3m.

4d Blue-brown, shaly clays with some *Liostrea* sp.

0.4m.

4c Impersistent, very thin shale band with ripple marked upper surface, burrows and serpulids on this surface. Unfossiliferous

0.03m.

4b Impersistent thin shelly limestone overlain by blue and brown clays of variable thickness with some *Liostrea* sp. and *Kallirhynchia* sp.

0.2m.

4a **Lower Nerinaea Bed**. Orange-brown sandy, soft oolite passing up into greenish grey, oolitic clay. *Aphanoptyxis eulimoides* (Lycett) prolific at top and frequent *Neridomus pseudocostata* (d'Orb), many smaller specimens (*N. cooksoni* Desl.) showing fine colour banding, *Liostrea hebridica* (Forbes), *Trigonia* sp., *Nucula* sp., frequent *Nuculana lachryma* (J.de C. Sow), *Pholadomya* and heterodont bivalve casts with *Kallirhynchia* sp. (crushed)

0.35m.

3. **CHIPPING NORTON LIMESTONE.** Massive, shelly, cross-bedded, oolitic limestone with intercalated softer, orange coloured, oolitic marls. Basal bed is more fossiliferous than the rest, with abundant *Liostrea* sp. and *Clypeus* sp., with echinoid spines, *Trigonia* sp., *Lima cardiformis* J. Sow., and *Strophodus* teeth. Finely preserved trace fossils (*Rhizocorallium)* also preserved

1.5-2m.

2. **HOOK NORTON LIMESTONE.** Massive, evenly-bedded, hard, oolitic limestone with softer intercalations, cross-bedded in places. Very unfossiliferous and only oyster shell debris and *Lima* cf. *cardiformis* J. Sow.

seen c. 10m.

Gap - strata obscured - junction not visible.

1. **CLYPEUS GRIT.** Rubbly, yellow, coarse oolitic (almost pisolitic) limestone with marl patches. *Parkinsonia* sp., *Cenoceras* cf. *inornatum* (d'Orb.), *Clypeus ploti* Salter, very abundant in nests but not as abundant as at Aston Bank and often overgrown by serpulids, *Holectypus* sp. frequent, *Stiphothyris tumida* (Davidson) - common in nests, *Rhacorhynchia regalis* S. Buckman - rare, with *Homomya* sp. and *Pholamya* sp. some in position of growth

seen 5m."

Base of Torren's (1969) section at Hornsleasow Quarry.

APPENDIX B2: Section at Hornsleasow Quarry described by author (*in* Callomon *et al*, 1993) with subsequent modifications made by the author regarding the position of the newly defined Charlbury Formation (Boneham & Wyatt, 1993) in the sequence

6. TAYNTON FORMATION. Shelly, oobiosparite limestone, occasionally flaggy and with intraclasts. Fauna includes bivalves (*Plagiostoma, Praeexogyra, Myophorella*), nautiloids (*Procymatoceras*) and rare *Procerites* ammonites. Becoming more fossiliferous toward base (partly obscured). Fragments of wood and fish teeth are also found.

partially overgrown, seen 2-3m

Gap (1m) - boundary obscured

5. CHARLBURY FORMATION.

5d Brown and green clays with poorly preserved large nerinid gastropods, *Aphanoptyxis* (Torren's Upper Nerinaea Bed)

formerly seen 0.45m

5c Limestone, very soft, oolitic, rubbly or flaggy. Numerous *Skolithos* burrows

partially obscured, seen 0.3m

5b Grey-green soft oolitic marl and brown oolitic clay. *Aphanopteryxis*, and oyster debris (Torren's Middle Nerinaea Bed)

partially overgrown, seen 0.45m

Top of section visible today

5a **Eyford Member.** Hard, fissile, flaggy, oolitic and arenaceous limestone, marly and rubbly at top, hard bed at base. Quasi-planar lamination and cross-bedding apparent, much shell-debris but few well preserved macrofossils (*Praeexogyra*); sharks teeth. (Channon's Bed 7 - "Stonesfield Slate")

2.5m

Sharp boundary

4. SHARPS HILL FORMATION.

4f. Clay, blue, shaly or marly, with casts of *Natica*-like gastropods, high-spired *Aphanoptxis*, the ryhnconellid *Kallirhynchia* and the bivalve *Modiolus imbricatus*, and much wood debris

0.5m

4e. **Coral Bed.** Much of this has been removed by quarrying and collectors, but well-preserved corals can still be found weathering out of the blue clay matrix (for faunal list see Torren's section)

partially obscured, 0.2-0.3m

4b-d. Shaly clays and thin limestones, blue to brown sparsely fossiliferous (Praeexogyra and Kallirhynchia) and somewhat overgrown section

0.5-0.6m

4a. **Nerinaea Bed.** (Lower Nerinaea Bed of Torren's) Soft, marly and oolitic clays, greenish-grey weathering at outcrop to orange-brown. Highly fossiliferous: Bivalves include *Praeexogyra hebridica, Pholadomya, Trigonia, Nuculana lachryma* and other biodebris, gastropods include *Aphanoptyxis*

eulimoides, *Neriodomus pseudocostata* and *N. cooksoni* (which exhibits colour banding), brachiopods, corals, crustacea and fish teeth

0.35m

Fairly sharp boundary

3. CHIPPING NORTON FORMATION. 'CHIPPING NORTON MEMBER'

3c-e. A sandwich of two resistant sets of cross-bedded oolites (c and e) separated by a recessive, thin bed of oolitic marls (d) forming the top of the main limestone sequence as seen in western and eastern quarry faces, with a platform on top (graphic log, Fig. 4.1). Bioclastic with abundant *Praeexogyra* which occasionally form encrusted surfaces, echinoid debris and fish teeth. Trace-fossils include *Rhizocorallium* and *Skolithos* burrows

2m

3b. **The Hornsleasow Clay Unit.** Impersistent, lenticular clays set within hollows in the oolites below. In thicker lenses two layers of clay could be discerned (section 4.2). Well-preserved terrestrial fauna, including small and large vertebrate remains, fresh-water gastropods, ostracods, coprolites and wood material (Chapter 5)

up to 0.8m

Undulating karstic surface, iron-stained and decalcified

3a. Massive, cross-bedded oolites, weathering brownish-grey. More fossiliferous than the higher beds, with abundant *Praeexogyra*, occasional *Clypeus*, echinoid spines and few other species of bivalve

1m

Well-defined boundary, marked by layer of marl and a slight change in facies, clearly recognisable in weathered faces of the western and eastern ends of the quarry, but not so obvious close-up in fresh sections. The beds below weather to a more golden-yellow colour.

2. CHIPPING NORTON FORMATION. 'HOOK NORTON MEMBER'. Massive, hard oolitic limestones in many well-bedded courses, which individually are often more or less cross-bedded. very sparsely fossiliferous, oyster debris being the main discernible macrofossil.

Total thickness now visible 7m

Base of quarry, lower beds formerly seen to south are now totally obscured. These include a further 3.8m of Hook Norton Limestone, and ~5m of Clypeus Grit (Bed 1, Torrens, 1969).

APPENDIX C1: Preparation of thin sections of limestone for petrographic description

Fifteen samples of rock were taken from the quarry section (Fig. C1). From these a small slab of rock, measuring no more than 30x30mm, of 1-2mm thickness was first cut with a diamond saw, then lapped and ground smooth on one side using successively finer grades of carborundum powder abrasive, 600F (100 micron), 800F (80 micron) and 1200F (50 micron). This side of the block is glued to a glass slide using epoxy resin or Lakeside cement, and then ground down to 0.03mm thickness using the carborundum abrasives. The rock section were sealed under a glass cover slip, using epoxy or Canada balsam glue and then examined at x40, x100 and x400 under transmitted plane polarised and cross-polarised light using a petrographic microscope (Prior model MP3500).

APPENDIX C2 : Preparation of stained acetate peels using the Dickson method of staining carbonate rocks

Ten hand specimens of limestone were taken from the field (Fig. C1). These were first cut in half with a diamond saw, and the cut surface lapped and buffed to smoothness using 600F and 800F carborundum powder. These samples were stained using the Dickson (1966) method of staining which is employed to distinguish different carbonate cements. Chemical staining works by producing a precipitate on the surface of a particular mineral. The staining solution prepared is a mixture in the proportions of 3:2 of two reagents that can differentiate between the ferroan phases within calcite and dolomite cements, they are Alizarin Red-S (0.2g/100ml of 1.5% hydrochloric acid, HCl) and potassium ferricyanide (2g/100ml of HCl).

The cut and polished surface of the rock was etched for 10 seconds in 1.5% HCl and then placed in the staining solution for about 60 seconds. The surface was washed with distilled water and allowed to air dry. The resulting colour stains using the combined solution for each type of carbonate cement usually found in limestones are:

Aragonite	Pink
Calcite	Pink
Ferroan calcite (>1% $FeCO_3$)	Mauve-Royal blue
Dolomite	Unstained
Ferroan dolomite (>1% $FeCO_3$)	Turquoise

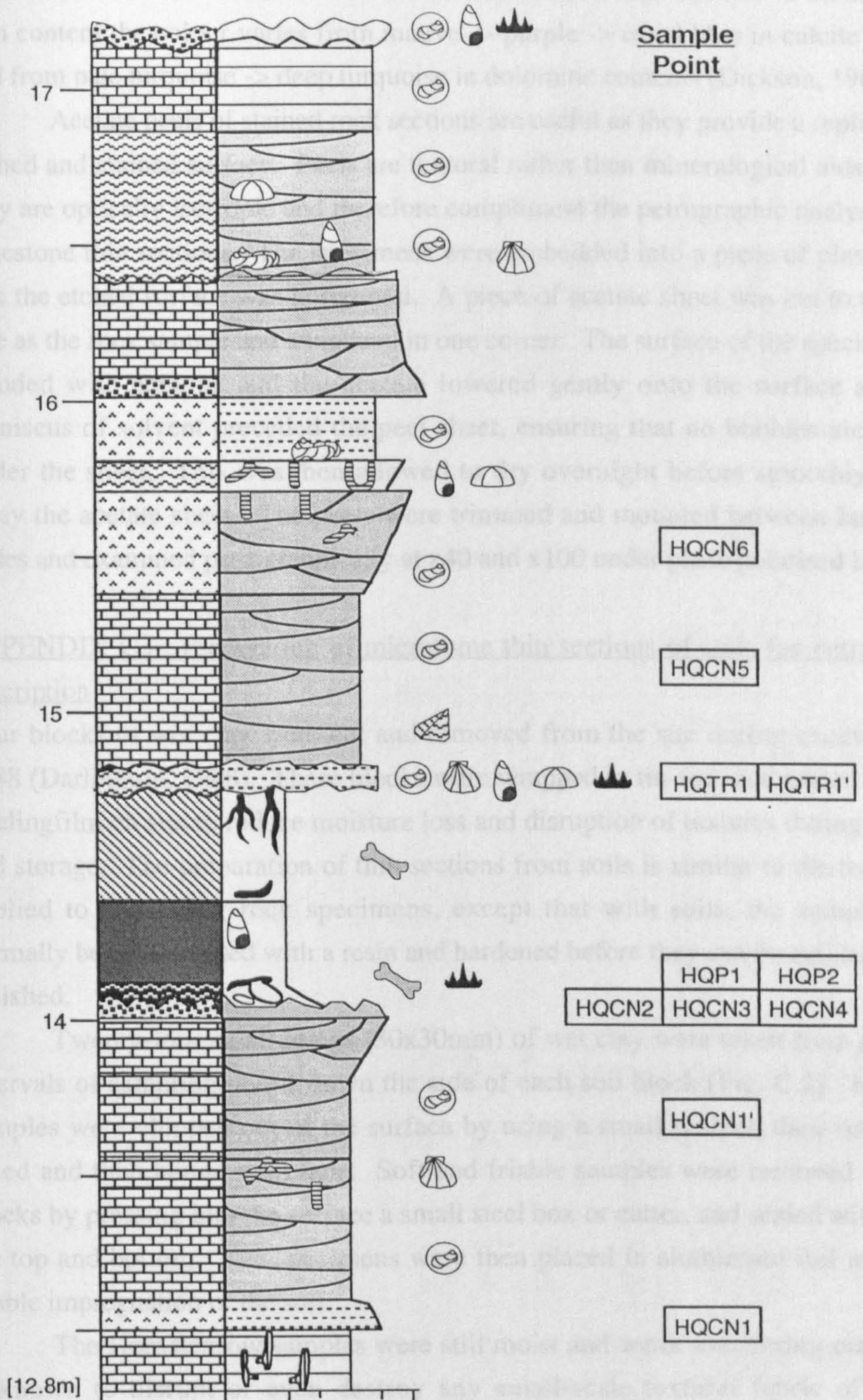


Figure C1. Graphic log of the Hornsleasow section showing position of sampling points

Potassium ferricyanide is very sensitive to the detection of iron in carbonates and will stain cements with as little as 1% iron carbonate within their matrix. With increasing iron content the colour varies from mauve -> purple -> royal blue in calcite cements and from pale turquoise -> deep turquoise in dolomitic cements (Dickson, 1966).

Acetate peels of stained rock sections are useful as they provide a replica of the etched and stained surface. Peels are textural rather than mineralogical aids because they are optically isotropic and therefore compliment the petrographic analysis of the limestone thin sections. The specimens were embedded into a piece of plasticine so that the etched surface was horizontal. A piece of acetate sheet was cut to the same size as the rock sample and annotated in one corner. The surface of the specimen was flooded with acetone and the acetate lowered gently onto the surface so that a meniscus of solvent preceded the peel sheet, ensuring that no bubbles are trapped under the sheet. This was then allowed to dry overnight before smoothly peeling away the acetate sheet. The peels were trimmed and mounted between large glass slides and examined petrographically at x40 and x100 under plane polarised light.

APPENDIX C3 : Preparation of microtome thin sections of soils for petrographic description

Four blocks of wet clay were cut and removed from the site during excavation, in 1988 (Darlington, 1988). These blocks were wrapped in tin-foil, and sealed in layers of clingfilm on site to reduce moisture loss and disruption of textures during removal and storage. The preparation of thin sections from soils is similar to the techniques applied to sectioning rock specimens, except that with soils, the samples must normally be impregnated with a resin and hardened before they can be cut, lapped and polished.

Twenty-four small lumps (30x30mm) of wet clay were taken from regularly intervals of sampling points down the side of each soil block (Fig. C.2). Hardened samples were chipped out of the surface by using a small scalpel, their orientation noted and then bound with tape. Soft and friable samples were removed from the blocks by pressing into the surface a small steel box or cutter, and sealed with tape at the top and bottom. The specimens were then placed in aluminium foil moulds to enable impregnation of the soil.

The Hornsleasow samples were still moist and water loss during preparation is known to disrupt or even destroy any small-scale textural fabric of the soil (FitzPatrick, 1984). Therefore, the water was slowly replaced by an increasing concentration of acetone in water over a period of a few weeks and then impregnated with dilute resin. The procedure for replacing soil moisture and impregnation of soil samples is described in minute detail in FitzPatrick (1984), and an outline is provided here. The specimens were covered to a depth of about 10mm with 10% acetone in

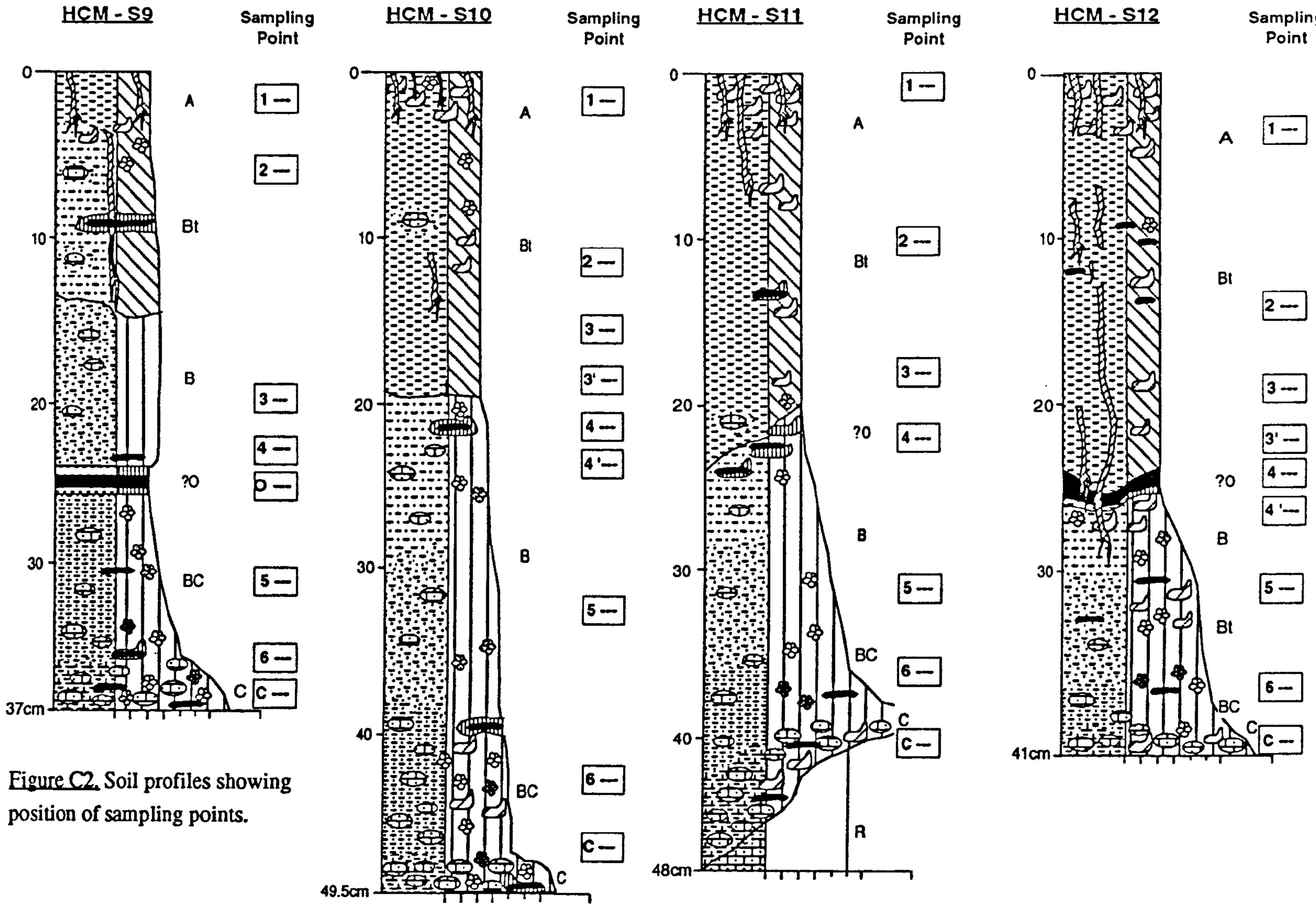


Figure C2. Soil profiles showing position of sampling points.

water and left to stand for 24 hours, in a closed vessel. Excess water-acetone mixture was removed by pipette, and the procedure was repeated nine times, increasing the acetone concentration by 10% at each successive stage. The procedure was repeated a further ten times using pure acetone to ensure that all the soil water had been exchanged for acetone.

Impregnation of the specimens can be considered the most crucial stage of the process, and can be achieved by using several different resins and methods (see FitzPatrick, 1984). Unfortunately, all but one of the resins tried and tested upon the Hornsleasow soil samples failed to impregnate (G. Webber, pers. comm., 1994). Generally, epoxy or polyester resins are used to impregnate soils, however we found that the epoxy resins (such as Araldite) reacted with the clays and caused them to swell considerably. Finally, the polyester resin Crystic B (FitzPtatrick, 1984) thinned in acetone was found to be the most successful (G. Webber, pers. comm., 1994). The procedure for impregnation of the soils containing acetone is as follows. Firstly the samples were covered to a depth of 5mm with the impregnating mixture, then a further 100ml is added and left overnight in a sealed container. The specimens were placed in a fume cupboard to allow the acetone to evaporate until the samples were covered to a depth of 5mm. A further 100ml of very dilute resin was then added and the specimens were left overnight as before. This procedure was repeated until the blocks had at least 10mm of resin above the level of the soil sample, which ensured that they were well impregnated and would not break apart upon cutting. The specimens were then left to stand for several days until weight loss was less than 0.1g per day and then placed in an oven at 25°C until further weight loss had ceased. The final procedure was to 'cure' the specimens at 40°C.

This procedure is very lengthy, but ensures that no cracks appear in the soils during preparation. The hardened samples were then removed from the foil moulds, cut, lapped and polished in the standard manner, before being attached to glass slides. These were then lapped and ground down to 24um thickness in the same way as for the limestone thin sections (Appendix C1), and a cover slip secured. Unfortunately, this procedure was experimental and some of the samples dried out too quickly during impregnation and fell apart, or disintegrated during grinding (G. Webber, pers. comm., 1994). However, of the original 24 samples, fifteen thin sections were made and described petrographically under plane and cross-polarised transmitted light at x40 and x100.

APPENDIX C4 : Description of the thin and polished sections and acetate peels made from the Hornsleasow limestones.

Each specimen was examined in three ways to determine the rock type:

(1) Hand specimens and polished blocks were examined in conjunction with stained acetate peels to facilitate recognition of the large scale textural features and general mineralogy of the rock type. Features looked for and recognised at this point include:

(a) Textural features such as lamination or bedding.

(b) Overall grainsize distribution and inclusion of any large distinct clasts or lenses of clasts. This would give an approximation of the type of limestone based upon a very simple classification scheme which splits carbonate rocks into calcirudite (grainsize > 2mm), calcarenite (grainsize between 2mm - 62um) and calcilutite (grainsize < 62um) (Tucker, 1981).

(c) Colour variation and weathering. Colour is described and also given a reference Munsell^R value, based upon the Munsell^R hue and chroma scales given for rocks (Rock-Color Chart Committee, 1991). Munsell^R values take the form of 5R or 10YR for 'hue', followed by a number for 'value' of light to dark on a decreasing scale from 8 - 2 and lastly a number for 'chroma', which is the degree of saturation of colour from neutral grey to purple (Rock-Color Chart Committee, 1991). For example a pale greyish orange-pink is described as 10R followed by 8 and then 2, so it's Munsell^R description is 10R 8/2.

(d) Approximate determination of the amount of cementation within the rock.

(2) Examination of thin section in natural daylight, in conjunction with acetate peels. This helps to recognise any distinct layering or lamination within the thin section before description is attempted. Also recognised at this point are the general grainsize distribution and sorting within the thin section and the inclusion of distinguishable zones of cementation.

(3) The thin section and acetate peels are then described fully under the petrographic microscope at x40, x100 and x400 magnification, in both plane and cross-polarised transmitted light. The classification of carbonate rocks involves the identification of clasts or 'allochems' and the estimation of the proportions of microcrystalline calcite or 'micrite' and sparry calcite or 'sparite' in the section (Adams *et al.*, 1984). Allochems are the large grains or clasts and include ooids, bioclasts, peloids and intraclasts. Micrite is the carbonate cement which is formed of crystals less than 5μm, hence microcrystalline, it is also known as 'carbonate mud' (Tucker, 1981). It can be introduced during deposition as a precipitate from sea-water or the disintegration of carbonate hard parts (e.g. the mictritic envelopes around shells caused by the boring of endolithic algae), or can be precipitated as a cement during early diagenesis (Adams *et al.*, 1984). Sparite is the coarse crystalline calcite (diameter over 5μm) which forms pore-filling cement and is usually a later diagenetic feature (Tucker, 1981). The two most useful classification schemes for carbonate

<table>
<tr><td colspan="6">Original components not organically bound together during deposition</td><td>Components organically bound during deposition</td></tr>
<tr><td colspan="3">of the allochems, less than 10% > 2mm diameter</td><td colspan="2">of the allochems more than 10% > 2mm</td><td rowspan="2"></td></tr>
<tr><td colspan="2">contains carbonate mud</td><td>mud absent</td><td>matrix supported</td><td>grain supported</td></tr>
<tr><td colspan="2">mud-supported</td><td rowspan="2">grain supported</td><td rowspan="2"></td><td rowspan="2"></td><td rowspan="2"></td></tr>
<tr><td>< 10% allochems</td><td>> 10% allochems</td></tr>
<tr><td>MUDSTONE</td><td>WACKESTONE</td><td>PACKSTONE</td><td>GRAINSTONE</td><td>FLOATSTONE</td><td>RUDSTONE</td><td>BOUNDSTONE</td></tr>
</table>

Table C1. Classification of limestones based upon depositional texture, after Dunham (1962). Rock names are in capitals.

Principal allochems in limestone			> 10% allochems		< 10% allochems		
			Sparite > micrite	Micrite > sparite	1-10% allochems		< 1% allochems
> 25% intraclasts			INTRASPARITE	INTRAMICRITE		Intraclasts INTRACLAST-BEARING MICRITE	
	> 25% ooids		OOSPARITE	OOMICRITE		Ooids OOID-BEARING MICRITE	
	< 25% ooids volume ratio bioclasts : peloids	>3:1	BIOSPARITE	BIOMICRITE		Bioclasts FOSSILIFER-OUS MICRITE	
		<1:3	PELSPARITE	PELMICRITE		Peloids PELOID-BEARING MICRITE	

Table C2. Classification of limestones based upon composition, after Folk (1959, 1962). Rock names are in capitals.

rocks are those proposed by Folk (1959, 1962) (Table C2) and Dunham (1962) (Table C1) each of the thin sections were described according to these schemes, based upon:

(a) Texture (Dunham, 1962). This is a useful scheme which classifies limestones on the basis of depositional texture, it is summarised in Table C1.

(b) Composition (Folk, 1959, 1962). The classification scheme of Folk divides carbonate rocks based upon the proportions of allochems, matrix (i.e. micrite) and cement (i.e. sparite) and hence, is a scheme based upon composition (Table C2).

Other features (Table C3) which give an indication of the depositional and diagenetic environment of the Chipping Norton Formation limestones at Hornsleasow were examined and described for each of the thin sections and these are listed below:

(c) Sorting. This gives an indication of the energy of the depositional environment and is based upon degrees of mechanical rounding of the allochems and the amount of carbonate mud present in the slide (Adams *et al.*, 1984). A rock containing over two-thirds micrite is derived from low-energy conditions, subequal amounts of spar and micrite within a section is known as 'poorly washed' and is considered to be of low-moderate energy, and rocks with over two-thirds spar cement are considered to be from moderate-high energy conditions. Within the latter, the size distribution and rounding of grains is also considered to also give an indication of sorting, with poorly sorted rocks containing angular fragments of allochems of differing sizes and well sorted rocks containing rounded and abraded particles of sub-equal size-range (Adams *et al.*, 1984).

(d) Cementation. The morphology and mineralogy of carbonate cement crystals within a limestone can give much information about the environment of cementation and diagenetic history of the rock. Four types of cement were noticed within the Hornsleasow sections, in variable proportions. Their generation is considered within section 3.4.2. They are :

(i) Pore-filling micrite or calcite microspar.

(ii) Radial-fibrous pore-lining calcite and syntaxial overgrowths of high magnesium calcite upon allochems

(iii) Drusy mosaic pore-filling ferroan calcite spar.

(iv) Equant low magnesium calcite vein-spar.

(e) Porosity. An evaluation of both type and amount of porosity within the limestone was examined. Porosity can be a primary depositional feature of the rock or may have formed from secondary dissolution as a result of diagenesis. There are three main types of porosity seen in limestones (Adams *et al.*, 1984), which are 'fabric-selective', 'non-fabric-selective' and 'partially-fabric-selective'. These are split up into several sub-classes, the most common of which are described below:

Over 2/3 micrite matrix				subequal spar & micrite	Over 2/3 sparite cement		
0-1% allochems	1-10% allochems	10-50% allochems	> 50% allochems		Sorting poor	Sorting good	Rounded & abraded
Micrite & dismicrite	Fossiliferous micrite	Sparse biomicrite	Packed biomicrite	Poorly-washed biosparite	Unsorted biosparite	Sorted biosparite	Rounded biosparite

Table C3. A range of textures shown in limestones (with Folk classified.rock types), after Adams *et al.* 1984

(i) Fabric-selective. Interparticle (matrix removed), intercrystal (cement removed), intraparticle (selective dissolution of parts of allochem), mouldic (allochems dissolved away), fenestral (dissolution along laminae or bedding planes), shelter (depositional hollows beneath curved allochems e.g. bivalve shells, or occur within chambered fossils e.g. gastropods; these are geopetal or way-up structures).

(ii) Non-fabric-selective. Fracture (cracks), channel (dissolution along cracks), vug (discrete dissolutional hollows).

(iii) Partially-fabric-selective. Breccia (physical breaking apart of rock may follow non-fabric selective cracks or fabric-selective fenestrae), boring (in shells is fabric-selective, otherwise is not), burrows, shrinkage (caused by desiccation of the sediment, these are also known as 'birdseyes'; Tucker, 1981).

(f) Grain diagenesis. The preservation and mineralogy of the grains within the limestones can also yield much information upon the original conditions of deposition and any post-depositional alteration and diagenesis of the rocks. This is particularly relevant for the Hornsleasow limestones which have undergone karstification and sub-aerial weathering in Bathonian times (section 4.4). The allochems within the Hornsleasow limestones show four types of mineralogical preservation or alteration, which are :

(i) Micritic envelopes and borings around the edges of allochems such as bioclasts and ooids caused by microbial activity. This is also known as micritisation.

(ii) Calcitisation or *in situ* replacement of aragonitic allochems often preserving textural details. Some well-preserved ooids exhibit strong compositional zoning with the outer lamellae being composed of ferroan calcite or even siderite.

(iii) Wholesale dissolution and replacement by ferroan drusy sparite.

(iv) Dissolution of carbonate cements (including siderite) to leave a residue of ferric hydroxide and aggregated iron oxides (see below).

The sections were not studied under cathode luminescence, because the high iron content within the ferroan calcite and siderite, and in the form of ferric oxyhydrites, inhibited the luminescence, meaning that high and low magnesium calcites were indistinguishable. This is a common phenomena with high ferroan carbonates (Tucker, 1981). Based upon all the above criteria, the carbonate rocks sectioned from the Chipping Norton Formation at Hornsleasow are described below:

(A) Unaltered Chipping Norton Formation limestones from below the karstic surface

HQCN1

(1) Shelly calcarenite with distinct compositional cross-laminations. Buff-coloured (10YR 8/2) weathering to a yellowish-grey (10YR 8/6). Shelly horizons are much better cemented than oolitic bands.

(2) Two thin sections cut.

HQCN1-1. Finely laminated (0.3-0.6mm) oolitic calcarenite taken from the basal part of sample HQCN1.

HQCN1-2. Shelly band, 10 mm in thickness, taken from near the top of sample HQCN1.

(3) Thin section descriptions.

HQCN1-1. (a) Grainstone, in places a packstone

(b) Modal composition (percentages) : Oopelsparite

Matrix (carbonate mud) : 10-15

Cement (sparite) : 25-30

Allochems : 55-65	Ooids : 20-25	Bioclasts : 10-12
	Peloids : 20-25	Intraclasts : 0
Other grains : <5	Quartz : 0	Opaques : 3-5

(c) Clasts rounded and abraded - rounded oosparite, in places subequal amounts of carbonate mud and sparite cement, poorly washed.

(d) (i) Micrite bridges between grains (may be depositional carbonate mud, which was partially lithified before sparite was cemented)

(ii) Rare pore-lining radial-fibrous calcite (stained pink)

(iii) Pore-filling drusy ferroan calcite (stained mauve)

(e) Very low interparticle, mouldic and intraparticle porosity, 2-5%

(f) (i) Micritic envelopes on 80-90% allochems and 40-45% wholly micritised

(ii) Calcitisation of 25-30% of allochems. 2-5% ooids show sideritic outer lamellae.

(iii) Rare wholesale or partial dissolution of bioclasts and replacement by ferroan drusy spar (5-10%)

HQCN1-2 (a) Rudstone, in places a floatstone

(b) Modal composition (percentages) : Biopelsparite

Matrix (carbonate mud) : 0-5 (up to 15% in floatstone)

Cement (sparite) : 50-60

Allochems : 40-45	Ooids : 5-10	Bioclasts : 10-15
	Peloids : 20-25	Intraclasts : 5-10

Other grains : <5 Quartz : 0 Opaques : 2-5

(c) Clasts rounded and abraded - rounded poorly washed biopelsparite

(d) (ii) Rare pore-lining radial-fibrous calcite (stained pink) and syntaxial overgrowths upon echinoid fragments

(iii) Coarse pore-filling drusy ferroan calcite (stained mauve)

(iv) Vug- and vein-filling large equant non-ferroan calcite

(e) Low interparticle porosity

(f) (i) Micritic envelopes on 80-90% allochems and wholly micritised grains 30-40%

(ii) Calcitisation of ooids

(iii) Wholesale dissolution of bioclasts and replacement by ferroan drusy spar (40-45%)

HQCN1'

(1) Well-cemented cross-bedded calcarenite. Large-scale calcitic veining and rock is riddled with cracks. Pale-cream (10YR 8/2) rock weathering to a warm yellowish buff colour (10YR 7/4).

(2) Thin section in daylight.

No structure discernible. Equigranular massive, well-cemented calcarenite. Unweathered.

(3) Thin section description.

(a) Grainstone

(b) Modal composition (percentages) : Biopelsparite

Matrix (carbonate mud) : 0-2

Cement (sparite) : 50-55

Allochems : 40-45 Ooids : 10-15 Bioclasts : 15-20

Peloids : 10-15 Intraclasts : <5

Other grains : <5 Quartz : (15-20 ooids) Opaques : < 2

(c) Clasts rounded and abraded, well sorted - rounded biopelsparite

(d) (ii) Extremely rare pore-lining radial-fibrous calcite within tiny pore spaces and syntaxial overgrowths

(iii) Pore-filling anhedral drusy ferroan calcite (stained mauve); within grains is not completely pore-filling

(iv) Sub-euhedral equant calcite spar, seen as pore-filling and within veinlets and vugs. Well cleaved and twinned low magnesium calcite

(e) Intraparticle, mouldic and interparticle porosity - partially infilled within ferroan calcite and later infilled within non-ferroan calcite. Also low channel and vug porosity

(f) (i) Micritic envelopes on 75-80% allochems, 45-50% allochems are wholly or partially micritised

(ii) Calcitisation of ooids and 1-2% bioclasts

(iii) Rare wholesale dissolution of bioclasts and incomplete replacement by ferroan drusy spar (25-30%)

(B) Unaltered Chipping Norton Formation limestones from above the clay lens

HQCN5

(1) Well-cemented (calcitic) cross-bedded calcarenite. Cream-grey (5Y8/1) rock weathering to a warm yellowish-buff colour (10YR 7/4).

(2) Thin section in daylight.

Planar laminated well-cemented calcarenite. Unweathered.

(3) Thin section description.

(a) Grainstone, in places packstone

(b) Modal composition (percentages) : packed oobiomicrite

Matrix (carbonate mud) : 20-25

Cement (sparite) : 10-15

Allochems : 55-60 Ooids : 20-25 Bioclasts : 15

Peloids : 10-15 Intraclasts : 5-10

Other grains : 5 Quartz : 3-5 Opaques : < 2

(c) Clasts rounded and well sorted - sorted packed oobiomicrite

(d) (i) Ooids are compositionally zoned - inner core calcitised green micrite -> outer margins ferroan rich micrite and siderite.

(iii) Pore-filling anhedral drusy ferroan calcite (stained mauve); within grains is not completely pore-filling

(e) <5% interparticle and mouldic porosity. Also low channel and vug porosity

(f) (i) Micritic envelopes on 80-90% allochems, including compositionally zoned ooliths. Partially and wholly micritised allochems make up 55-60%.

(ii) Calcitisation of ooids and 2-5% bioclasts. Ooids show well-preserved concentric zoning and radial structure (in cross polars) about 15-20% show sideritic outer lamellae

(iii) Very rare wholesale dissolution of bioclasts and incomplete replacement by ferroan drusy spar (5-10%)

HQCN6

(1) Coarsely oolitic, shell fragmental limestone (calcarenite). Friable and not well-cemented. Creamy-buff (10YR7/4) rock weathering to a warm dark yellowish-buff colour (10YR 6/6).

(2) Thin section in daylight.

Muddy matrix, laminated limestone with rare allochems.

(3) Thin section description.

(a) Wackestone

(b) Modal composition (percentages) : sparse oopelmicrite

Matrix (carbonate mud) : 40-50

Cement (sparite) : 5

Allochems : 40-45	Ooids : 15-20	Bioclasts : 5-10
	Peloids : 10-15	Intraclasts : 2-5
Other grains : <5	Quartz : <2	Opaques : < 2

(c) Clasts rounded, however poorly sorted - unsorted oopelmicrite

(d) (i) Ooliths are compositionally zoned - inner core calcitised green micrite -> outer margins ferroan rich micrite and siderite. Carbonate mud calcitised to form pore-filling cement

(e) Low intraparticle and intercrystal porosity, <5%; channel and fracture porosity may occur during preparation of thin section

(f) (i) Micritic envelopes on >95% allochems, including compositionally zoned ooliths. 65-70% micritised allochems.

(ii) Calcitisation of ooids and 2-5% bioclasts. Ooids show well-preserved concentric zoning and radial structure (in cross polars) and also aggregated ooids formed during reworking. Most ooids show sideritic outer lamellae.

(iii) Very rare wholesale dissolution of bioclasts and incomplete replacement by ferroan drusy spar (<5%)

(C) The 'transgression layer' sandy limestones

HQTR1

(1) Pale creamy-buff (10YR 8/2) sandy limestone, which is distinctly planar and cross laminated. There are thin lenticular shelly bands containing whole and broken biodebris, including oysters, other bivalves, brachiopods, echinoid debris and fish teeth. Weathers to a pale yellow-brown (10YR 6/2) friable sand. Well cemented, although decalcified along bedding planes (fenestral porosity).

(2) Three thin sections cut.

HQTR1-1. Finely laminated (1.0-5.0mm scale) oolitic calcarenite taken from the basal part of sample HQTR1.

HQTR1-2. Better cemented, but not as clearly laminated oolitic calcarenite taken from the mid part of sample HQTR1

HQTR1-3. Finely laminated (0.5-2.0mm scale) calcarenite, with shelly bands and decalcified cracks following bedding planes. Taken from the top part of the sample.

(3) Thin section descriptions.

HQTR1-1. (a) Grainstone, in places a packstone

(b) Modal composition (percentages) : Oosparite, micro-oolitic

Matrix (carbonate mud) : 15-20

Cement (sparite) : 15-20

Allochems : 50-60	Ooids : 25-30 (<40)	Bioclasts : 10-15
	Peloids : 10-15	Intraclasts : 0
Other grains : 5-10	Quartz : 5-10	Opaques : 0-2

(c) Clasts rounded and abraded - rounded oosparite, in places subequal amounts of carbonate mud and sparite cement, poorly washed.

(d) (iii) Pore-filling drusy slightly ferroan calcite (stained mauve-pink)

(e) Low interparticle, mouldic and intercrystal porosity, <10%

(f) (i) Micritic envelopes on 90% allochems and 50-55% allochems partially or wholly micritised.

(ii) Calcitisation of ooids and 2-5% bioclasts. Most ooids show compositional zoning from inner micrite to outer ferroan-micrite and siderite.

(iii) Very rare wholesale dissolution of bioclasts and replacement by ferroan drusy spar (5-10%)

HQTR1-2 (a) Grainstone

(b) Modal composition (percentages) : Oosparite, micro-oolitic

Matrix (carbonate mud) : 5-10

Cement (sparite) : 20-25

Allochems : 50-55 Ooids : 25-30 Bioclasts : 10-15

Peloids : 10-15 Intraclasts : 2-5

Other grains : 10-15 Quartz : 5-10 Opaques : 2-5

(c) Clasts rounded and abraded, however larger bioclasts also present - poorly sorted oosparite

(d) (ii) Rare calcitic syntaxial overgrowths upon echinoid fragments (<5%)

(iii) Pore-filling drusy slightly ferroan calcite (stained mauve-pink)

(e) Low interparticle and intercrystal porosity, 2-5%

(f) (i) Micritic envelopes on 90-95% allochems. 55-60% allochems are partially or wholly micritised.

(ii) Calcitisation of ooids and 2-5% bioclasts; ooids show well-preserved radial structures and concentric lamellae, with zoning from inner micrite to outer ferroan-micrite and siderite.

(iii) Rare wholesale dissolution of bioclasts and replacement by ferroan drusy spar (5-10%)

(iv) Detrital quartz is fairly common, and is subrounded to rounded. It is largely unaltered by weathering (weathering stage 0(1); see next section). Minimal dissolution of carbonate cements (including siderite) to leave a residue of ferric hydroxide and aggregated iron oxides.

HQTR1-3 (a) Grainstone

(b) Modal composition (percentages) : Oopelsparite, micro-oolitic

Matrix (carbonate mud) : 5-10

Cement (sparite) : 20-25

Allochems : 55-65 Ooids : 25-30 Bioclasts : 10-15

Peloids : 15-20 Intraclasts : 5-10

Other grains : 5-10 Quartz : 5-10 Opaques : <2

(c) Clasts rounded, some larger bioclasts-- poorly sorted biooosparite

(d) (i) Micrite bridges between grains (may be depositional carbonate mud, which was partially lithified before sparite was cemented)

(iii) Pore-filling drusy slightly ferroan calcite (stained mauve-pink)

(e) Low interparticle and channel porosity, 5-10%

(f) (i) Micritic envelopes on 90-95% allochems. 65-70% allochems are micritised.

(ii) Calcitisation of ooids and 5-10% bioclasts; ooids show well-preserved radial structures and concentric lamellae, compositionally zoned from inner micrite to outer ferroan-micrite and sideritic lamellae. Also aggregated ooids and superficial ooids present

(iii) Very rare wholesale dissolution of bioclasts and replacement by ferroan drusy spar (<5%)

(iv) Detrital quartz is fairly common, and is subangular to subrounded. It is unaltered by weathering (weathering stage 0; see next section). Rare dissolution of carbonate cements (including siderite) to leave a residue of ferric hydroxide and aggregated iron oxides.

HQTR1'

(1) Sandy and friable buff-orange (10YR 7/4) calcilutite. Contains large rip-up clasts of grey-white (5Y 8/1) oolitic calcarenite and large shell fragments. Shows fine laminations and is weathered along these bedding planes (fenestral porosity). Weathers to a warm yellow-brown (10YR 6/6).

(2) Thin section in natural daylight

Large (0.5-2.0mm) sub-rounded clasts of oolite set within a green-brown carbonate mud matrix.

(3) Matrix:

(a) Wackestone

(b) Modal composition (percentages) : Oomicrite

Matrix (carbonate mud) : 45-50

Cement (sparite) : 2-5

Allochems : 35-45-	Ooids : 15-20	Bioclasts : 5-10
	Peloids : 10-15	Intraclasts : 5-10
Other grains : 10-15	Quartz : 5-10	Opaques : 2-5

(c) Clasts rounded and well sorted - packed oomicrite

(d) (iii) Vug-filling drusy ferroan calcite (stained mauve)

(e) Vug and fenestral porosity (birdseye structures) 5-10%; and rare mouldic porosity

(f) (i) Micritic envelopes on 90-95% allochems. 60-65% allochems show partial or whole micritisation.

(ii) Calcitisation of ooids and 5-10% bioclasts. Rare compositionally zoned ooids (1-2%) which show inner micritic, outer ferroan micritic and sideritic lamellae.

(iii) Very rare wholesale dissolution of bioclasts and replacement by ferroan drusy spar (<5%), usually associated with disaggregated oolite intraclasts.

(iv) Dissolution of carbonate cements (including siderite) to leave a residue of ferric hydroxide and aggregated iron oxides

Intraclasts:

(a) Grainstone

(b) Modal composition (percentages) : Oobiosparite

Matrix (carbonate mud) : 0

Cement (sparite) : 45-55

Allochems : 35-45	Ooids : 10-15	Bioclasts : 10-15
	Peloids : 5-10	Intraclasts : 2-5
Other grains : <10	Quartz : <5	Opaques : <2

(c) Clasts rounded and well sorted - rounded biooosparite

(d) (ii) Rare radial-fibrous calcitic coatings upon grains and syntaxial overgrowths upon echinoid fragments (<5%)

(iii) Pore-filling drusy ferroan calcite (stained mauve)

(e) Very little interparticle and fracture porosity (<2%)

(f) (i) Micritic envelopes on 90-95% allochems. 50-55% allochems micritised.

(ii) Calcitisation of ooids and 1-2% bioclasts

(iii) Rare wholesale dissolution of bioclasts and replacement by ferroan drusy spar (10-15%)

APPENDIX C5 : Description of the Hornsleasow paleosols in thin section

Each thin section was studied in two ways to determine the features of the soil:

(1) Examination of the thin section in natural daylight. This helps to recognise any distinct colour mottling or banding and areas of distinct weathering or alteration within the thin section before description is attempted. Also recognised at this point are the general grainsize distribution and sorting within the thin section and the inclusion of distinguishable zones of cementation and impregnation, and any significant grain-loss during preparation.

(2) Examination and description of the thin sections under the petrographic microscope at x40, x100 and x400 magnification, in both plane and cross-polarised transmitted light. The following features were looked for and examined in the soil sections:

(a) Structure. The structure refers to the total organisation and spatial distribution of the soil thin section (FitzPatrick, 1984). It largely depends upon the degree and type of aggregates or 'peds' within the soil and the distribution or pores

and pore space. The most commonly observed structures in the Hornsleasow thin sections are: 'angular blocky' (peds broken up into angular pieces, with low porosity); 'bridge' (fine material forming coatings and bridges between grains, peds are not formed, this is seen in the more arenaceous-rich soils and karstic limestones); 'composite' (a combination of two or more structural types); 'crumb' (peds broken up into irregular shaped crumbs which are totally separated by the intervening pore-space); 'granular' (sub-spherical peds or grains and frequent pore-space); 'irregular blocky' (peds broken up into irregular pieces, with low porosity); 'laminar' (elongated and aligned peds, oriented with bedding); 'massive' (continuous soil phase, peds rare or absent).

(b) Pores and porosity. Pores are either continuous or discrete features of the soil structure, which break the soil up into peds.

(c) Passages. These include those passages made by burrowing soil organisms (i.e. trace fossils) and those made by rootlets.

(d) Organic matter. This includes structured and unstructured plant debris (section 5.3.1), and faecal material.

(e) Rock fragments and detrital grains. A rock fragment is a soil clast which contains two or more minerals, whereas a detrital grain is an individual mineral grain, both are derived from the parent rock. They are also known as 'skeleton grains' (Retallack, 1990)

(i) Distribution and texture. There are four grain fabrics (Retallack, 1990) : 'granular' (grains in point-to-point contact, with little or no intervening fine-grained material); 'intertextic' (grains in contact, but with occasional intergranular fine-grained material); 'agglomeroplasmic' (fine-grained material forms local pockets and an incomplete matrix to the skeleton grains); 'porphyroskelic (skeleton grains are set within a finer matrix)

(ii) Grainsize and sorting. The detrital grains range from fine sand to very coarse sand, and from well-sorted to very poorly sorted.

(iii) Composition and proportion of rock (in modal percentages). The detrital grains within the Hornsleasow soils are mainly composed of fragmental and dissolved Chipping Norton Formation limestones, and include oolite fragments, disseminated ooids, peloids and bioclasts, and quartz grains which may be derived from the interior of ooids or which are terrigenous in origin. The presence of 'shattered quartz' or that derived from the weathering of volcanic ash, was also noted here (see Chapter 3).

(f) Matrix. This is the fine-grained material surrounding the grains

(i) Colour in plane-polarised light. This property depend upon the presence of different colour-bearing elements such as iron and manganese, their oxidation states and the presence of organic matter. Matrices often show colour

variation and patterns which result from localised removal or concentration of material. These colour variations are known as 'mottling' when the various regions have distinct boundaries and 'marbling' when they are diffuse (FitzPatrick, 1993). Colours most noticeable in the Hornsleasow soils are: opaque matrices (indicate large amounts of iron and manganese oxides, may be deep red if haematitic or black if pyrolusite is dominant, common in lateritic soils); reddish-brown to brownish yellow matrices (indicate presence of goethite in various states of hydration, often speckled due to presence of tiny crystals of goethite 1-2um, these are common in tropical soils); brown to dark-brown matrices (indicate the presence of ferrihydrites and colloidal organic matter, may be isotropic in cross polarised light); olive and olive-grey matrices (indicate iron is present in the ferrous state and are common in horizons permanently saturated with water); pale brown to pale yellow matrices (indicate leaching of coloured compounds).

(ii) Microfabric. Clay matrices viewed under cross-polarised light can either be optically isotropic or anisotropic (FitzPatrick, 1984, 1993). Anisotropic fabrics or 'bright clay' are caused by the alignment and organisation of clay particles within the soil as determined by soil-processes, such as wetting and drying, whilst isotropic fabrics contain randomly orientated particles (Retallack, 1990). The interference colours of these fabrics are usually first order greys-oranges, depending upon mineralogy of the clays and iron content. Orange and red birefringence indicates high iron contents. White and grey birefringence indicates the presence of kaolinite and strong yellows and orange indicates the presence of 2:1 clay minerals, such as smectite (FitzPatrick, 1993).

(iii) Zones of anisotropy. Depending upon the degree of anisotropy, the anisotropic components of the matrix can show distinct patterns (FitzPatrick, 1984, 1993), these are : 'anisotropic flecks' (random patterns of discrete anisotropic regions or domains); 'anisotropic zones' (zones of optical anisotropism which run through the matrix); 'anisotropic aureoles' (occur around grains, pores and rock fragments); 'anisotropic lines' (thin planes of anisotropic material occurring along shear-planes). Anisotropic zones can be described as : 'continuous' (anisotropic domains which have coalesced and may surround a ped or pore); 'fibrous' (clusters of long domains with parallel alignment, may conform to the surface of a ped or pore and may occur singly or in multiple zones); 'striated' or 'streaky' (composed of short domains with parallel alignment, these have diffuse borders and give an overall streaky appearence, mainly randomly orientated within a section). There are many more terms for optical anisotropic matrices (FitzPatrick, 1984), but most of these can only be determined under circularly polarised light. Also other authors list differing terms and descriptions for microfabrics (e.g. Retallack,

1990, Bullock *et al.*, 1985), these are not included in this description for clarity, but are discussed in the main text.

(g) Grain and aggregate coatings. These are usually clay or ferric oxide/hydroxide coatings. They also include the anisotropic aureoles of fine-clay matrix.

(h) Secondary or pedogenic minerals. A mineral or concretion which has formed *in situ* within the soil, the commonest in the Hornsleasow paleosols are iron oxides (goethite, hematite) and limonite, calcite and phosphate.

(i) Weathering of clasts. This is based upon skeleton grains, such limestone clasts, allochems and quartz grains. Quartz in particular yields the most information upon chemical weathering within the soils, as this will give the top value for degree of alteration. There are five classes of alteration for quartz grains (FitzPatrick, 1984):

Class 0 : Fresh; <2.5% of mineral altered

Class 1 : Slightly altered; 2.5-25% of mineral altered and dissolved

Class 2 : Moderately altered; 25-75% of mineral altered and dissolved

Class 3 : Strongly altered; 75-97.5% of mineral dissolved away

Class 4 : Completely altered; >97.5% of mineral dissolved away

Based upon all the above criteria, the paleosols sectioned from the clay lens at Hornsleasow are described below:

(A) Green Clay - 'A' horizon

HQS9-1

(1) Poorly impregnated section, showing only minute traces of clay, which are badly cracked up. No structure discernible.

(2) (a) Massive structure

(b) Discrete irregular and sinuous pores

(c) Root-hair passages. These are tiny (0.01mm width, 1mm length) sinuous cross-cutting features, which are coated with limonite haloes.

(d) Rootlets (limonitised), very rare structured cuticle or 'cutin' and decomposed, unstructured carbonaceous fragments or 'inertinite' associated with limonite stains (<5% of rock).

(e) (i) Agglomeroplasmic grain fabric

(ii) Fine to medium sand grains and silt, moderately sorted.

(iii) Composition: Oolite : 0 Ooids : 1-3

Bioclasts : 5-10 Peloids : 2-5

Quartz : 15-20 (2-5 shattered quartz)

(f) (i) Pale speckled buff-brown, mottled toward darker golden-brown mottles around organics.

(ii) Moderately anisotropic microfabric. Bright clay shows 1st order interference greys and yellows. Mottles near organics are reds and oranges.

(iii) Streaky or striated anisotropy in places zones become more fibrous and may conform to ped and grain surfaces. Weakly developed anisotropic aureoles are developed around quartz grains, allochems and organics. In areas with few grains anisotropic flecks are seen.

(g) 40% of allochems show fine limonitised coatings or strong weathering rinds with inclusions, which are reddish-brown granules of goethite and red-black hematite. Organics have iron-rich bright clay aureoles.

(h) Goethite, limonite and some hematite.

(i) Mixed weathering. Allochems badly altered to ferruginous opaques and unstructured brown-green ?micritic residue, particularly around the edges of grains and in the outer lamellae of ooids. Some calcitised bioclasts (especially echinoderm fragments) are better preserved but show dissolution at edges and large opaque-rich weathering rinds. Ooid and detrital quartz grains exhibit cracking and pitting, stage 0(1). Shattered quartz and feldspar are fairly fresh, show some pitting, stage 0.

HQS10-1

(1) Well impregnated section, with much oolitic detritus and organics. Shows weak lamination and distinct yellow and brown mottles.

(2) (a) Massive structure.

(b) Sinuous, continuous pores associated with the massive structure, elsewhere pores are more irregular and discrete. Numerous bifurcating sinuous pores.

(c) Root-hair passages. These are tiny sinuous cross-cutting features, which are coated with limonite haloes. Numerous, make up to 1-2% of section.

(d) Rootlets (limonitised), discrete fragments of cutin and inertinite associated with limonite stains or haloes (<5% of rock).

(e) (i) Agglomeroplasmic to intertextic grain fabric

(ii) Fine to very coarse sand grains, very poorly sorted.

(iii) Composition: Oolite : 5-10 Ooids : 5-10
Bioclasts : 15-20 Peloids : 3-5
Quartz : 15-20 (3-5 shattered quartz)

(f) (i) Pale speckled yellow-buff, mottled red-brown toward organic material and slightly khaki-brown in other places.

(ii) Moderately anisotropic microfabric. Bright clay shows 1st order interference colours, including yellows and oranges, and reds near organics and allochems.

(iii) Discontinuous striated and streaky bright clay zones, with few well-developed fibrous zones. Poorly developed anisotropic microfabric aureoles developed around quartz grains only.

(g) Bright clay coatings around small grains, pores surround larger grains and oolite fragments. Also ferruginous granules surround allochems and coalesce in places to form a thick coating around the weathering rind. These are also associated with rootlets and some pore-spaces.

(h) Goethite and limonite

(i) Mixed weathering. Ooids badly decalcified leaving ferruginous opaques in the outer lamellae and inclusions of within inner lamellae. Calcitised bioclasts are better preserved but show dissolution to a dark-grey ?micritic residue followed by decalcification around the edge to form thick weathering rinds of brown limonite and granular reddish goethite. Ooid and detrital quartz grains exhibit cracking and pitting, stage 0(1). Shattered quartz are clear, stage 0.

HQS12-1

(1) Impregnated large fragment of greenish-brown clay. Little to no detrital material. Shows large scale cracking and loss of matrix during grinding.

(2) (a) Composite irregular-blocky and angular-blocky structure, massive elsewhere.

(b) Frequent pore-space around peds with irregular and sinuous pores, discrete cracking pore-space within peds

(c) Root-hair passages. These are tiny sinuous cross-cutting features, which are coated with limonite-haloes and contain black specks of carbonaceous material in their core or voids.

(d) Structured carbonaceous rootlets (limonitised), and decomposed, unstructured 'inertinite' associated with limonite stains (<5% of rock) or yellow-brown mottles.

(e) (i) Porphyroskelic grain fabric

(ii) Fine sand grains and silt, very well sorted.

(iii) Composition: Oolite : 0 Ooids : 0
Bioclasts : 0 Peloids : 0
Quartz : 5-10 (no shattered quartz)

(f) (i) Speckled buff-khaki brown, with warm orange-brown mottles associated with organics and discrete aggregates.

(ii) Strongly anisotropic microfabric. Bright clay shows 1st order interference colours, including greys and oranges in anisotropic zones, pale yellows and greys around pores and through anisotropic lines and reds and oranges in orange-brown mottles near grains and organics.

(iii) Striated and streaky bright clay zones. In red-brown mottles clay is tightly woven to produce continuous zones. Anisotropic lines cross through earlier fabrics.

(g) Iron-rich mottles surround organics.

(h) Microgranules of red goethite and ?hematite give grey its speckled appearence.

(i) Quartz grains exhibit cracking and pitting, and some weathering rinds have developed; stage 1.

(B) Green Clay - 'Bt' and 'B' horizons

HQS9-2

(1) Two thin sections cut:

HQS9-2(1). Well impregnated section, with much oolitic detritus and opaques

HQS9-2(2). Well impregnated section, with much oolitic detritus and opaques. Some grain and matrix loss during preparation, especially around the edges of the section.

(2) HQS9-2(1).

(a) Massive

(b) Rare pore-space, pores discrete and irregular

(c) Rare root-hair passages. These are tiny sinuous cross-cutting features, which are coated with limonite-haloes and contain black specks of carbonaceous material in their core.

(d) Structured carbonaceous rootlets (limonitised), and decomposed, unstructured 'inertinite' associated with limonite stains (5-10% of rock).

(e) (i) Agglomeroplasmic to intertextic grain fabric (especially near oolite fragments).

(ii) Fine - coarse sand grains, poorly sorted.

(iii) Composition: Oolite : 3-5 Ooids : 5-10
Bioclasts : 10-15 Peloids : 2-3
Quartz : 15-20 (3-5 shattered quartz)

(f) (i) Speckled buff-khaki brown, with warm orange-brown mottles around grains

(ii) Weakly anisotropic microfabric. Bright clay shows 1st order interference colours, including reds and oranges in orange-brown mottles.

(iii) Striated and streaky bright clay microfabric zones, surrounding areas of non-orientated flecked anisotropic clay. In red-brown mottles clay occurs as continuous zones or anisotropic aureoles around allochems.

(g) Iron-rich mottles surround organics and allochems. Ferric oxides and hydroxides surround rootlets and impregnate the outer surface of rotten allochem weathering rinds to form an incomplete opaque coating.

(h) Goethite and limonite. Small grains of calcite.

(i) Ooids show weathering rinds and outer lamellae are decalcified. Calcitised bioclasts show large dark brown weathering rinds with outer decalcified opaque-rich coatings, but are generally in better condition. Oolite clasts show dissolution of sparite cement and grain-loss. Detrital and ooid quartz grains exhibit cracking and pitting; stage 0(1). Shattered quartz is clear or slightly pitted, stage 0.

HQS9-2(2).

(a) Massive

(b) Rare pore-space, pores discrete and sinuous, some bifurcate

(c) Root-hair passages. Tiny, cross-cutting features with limonite haloes which contain either structured cuticle (red) or black specks of inertinite.

(d) Decomposed, unstructured 'inertinite' specks (<5% of rock) and rare structured cutin fragments and rootlets.

(e) (i) Agglomeroplasmic grain fabric

(ii) Fine - very coarse sand grains, very poorly sorted.

(iii) Composition: Oolite : 5-10 Ooids : 3-5
Bioclasts : 5-10 Peloids : 3-5
Quartz : 10-15 (2-5 shattered quartz)

(f) (i) Speckled buff-khaki brown, with few darker brown and orange-brown mottles around allochems and organics.

(ii) Moderately anisotropic microfabric. Bright clay shows 1st order interference colours which are mostly greys and yellows. However, in orange-brown mottles around organics they include reds and oranges.

(iii) Streaky and striated bright clay zones, surrounding areas of non-orientated flecked anisotropic clay. Weakly-developed anisotropic aureoles. Continuous and fibrous zones also conform to ped and grain surfaces.

(g) Iron-rich mottles surround organics and many allochem grains. Rare ferric oxides and hydroxides surround some allochems and rootlets. Goethite granules impregnate surface of weathering rinds to form incomplete opaque coatings.

(h) Goethite and limonite. Interparticle fibrous calcite microspar coatings.

(i) Ooids and allochems show weathering rinds. Outer lamellae of ooids are altered to ?micritic residue and decalcified to leave an opaque residual coating of limonite/goethite. Calcitised bioclasts show weathering rinds, but are in better condition. Oolite clasts show dissolution of sparite cement along cracks and upon the outer edges to a brown ?micritic weathering rind. Ooid and detrital quartz grains exhibit cracking and pitting; stage 0(1). Shattered quartz crystals are fresh, stage 0.

HQS9-3

(1) Poorly impregnated specimen, with only a small amount of over-ground sectioned material left upon the slide. Matrix supported with few detrital grains, no other structures discernible

(2) (a) Massive.

(b) Frequent pore-space, connected or discrete sinuous pores, some bifurcate

(c) Rare root-hair passages. These are tiny (0.01mm width, 1mm length) sinuous cross-cutting features, which are coated with limonite-haloes and contain structured cuticle and are infilled with calcite.

(d) Structured cutin rootlets (limonitised), and decomposed, unstructured 'inertinite' associated with limonite stains (2-3% of rock).

(e) (i) Agglomeroplasmic-porphyroskelic grain fabric

(ii) Fine - very coarse sand grains, very poorly sorted.

(iii) Composition: Oolite : 0 Ooids : 2-5

Bioclasts : 5-7 Peloids : 1-2

Quartz : 10-15 (5-7 shattered quartz)

(f) (i) Speckled buff-pale brown, in places greenish-brown, and small-scale warm orange-brown mottles associated with organics.

(ii) Weakly anisotropic microfabric. Bright clay shows 1st order interference colours, including pale greys and yellows, and reds and oranges in orange-brown mottles and anisotropic aureoles.

(iii) Streaky and fibrous bright clay surrounding areas of non-orientated flecked anisotropic clay. Well-formed anisotropic aureoles.

(g) Iron-rich mottles surround organics. Ferric oxides and hydroxides surround rootlets. Granules of ferric oxides surround some ooids and impregnate weathering rinds.

(h) Goethite and hematite granules and limonite stains. Nodular equant calcite surround some grains and infills rootlets.

(i) Ooids and peloids show weathering rinds and outer lamellae of ooids are decalcified. Calcitised bioclasts show strong weathering rinds, with opaque

inclusions and coatings, but are in better condition. Detrital and ooid quartz grains exhibit cracking and pitting; stage 0(1). Shattered quartz is clear, stage 0.

HQS10-3

(1) Not well impregnated, only a small amount of material left after sectioning. Matrix supported, with very few detrital grains.

(2) (a) Massive, in places developing a composite crumb-laminar structure

(b) Numerous pores, mostly continuous and irregular

(c) Rare root-hair passages. Possible cross-section of limonitised rootlet, with concentric lamellae (may also be a limonite nodule) and fine long sinuous root passages. These have an outer limonitised halo, contain carbonaceous material and are infilled with calcite.

(d) Limonite rootlets; opaques associated with organic material and unstructured inertinite 1-2%

(e) (i) Agglomeroplasmic-porphyroskelic grain fabric

(ii) Fine - medium sand grains, moderately sorted.

(iii) Composition: Oolite : 0 Ooids : 5-7

Bioclasts : 5-7 Peloids : 2-5

Quartz : 10-15 (5-10 shattered quartz)

(f) (i) Pale buff-khaki, with faint orange-brown mottles around organics

(ii) Well-formed anisotropic microfabric. Bright clay shows 1st order interference greys and yellows.

(iii) Streaky and continuous bright clay zones surrounding peds. Interior of peds a woven-like flecked bright clay. Anisotropic aureoles well developed.

(g) Iron-rich bright clay mottles surround organics and some allochems. Nodular ferric oxides and hydroxides surround some allochems, forming outer coatings to weathering rinds.

(h) Goethite granules and limonite stains.

(i) Ooids and bioclasts show strong weathering rinds, but the interiors are in quite good condition, i.e. inner lamellae are still distinguishable within ooids, outer lamellae contain opaque inclusions. Ooid and detrital quartz grains exhibit cracking and pitting; stage 0(1). Shattered quartz exhibit some pitting, stage 0(1).

HQS11-2

(1) Fairly successful impregnation of fine-grained matrix, although some has been lost during grinding. Extremely fine-grained soil, little or no detrital grains. Large scale yellow and brown mottling and presence of organics noticed.

(2) (a) Massive in places angular-blocky structure

(b) Numerous pores, discrete and irregular. Some bifurcate.

(c) None recognised

(d) Structureless inertinite associated with limonitised aureoles

(e) (i) Porphyroskelic grain fabric

(ii) Fine sand grains, well sorted.

(iii) Composition: Oolite : 0 Ooids : 0

Bioclasts : 0 Peloids : 0

Quartz : 5-10 (1-2 shattered quartz)

(f) (i) Buff-khaki, with dark brown and pale golden-yellow mottles around organics

(ii) Moderate anisotropic microfabric. Bright clay shows 1st order interference colours. Orange and reds near organics.

(iii) Streaky and fibrous anisotropic zones surround peds with flecked or woven bright clay. Anisotropic aureoles well developed around quartz grains. Anisotropic lines cross other zones of anisotropy.

(g) Iron-rich mottles surround organics.

(h) Goethite granules and limonite.

(i) Detrital, shattered and ooid quartz grains exhibit some pitting; stage 0.

HQS12-2

(1) Very poor section, poorly impregnated and hardly any matrix left upon the slide. Structure is indiscernible.

(2) (a) Irregular-blocky structure

(b) Numerous pore space, continuous and irregular around peds, discrete within peds.

(c) Not recognised.

(d) Unstructured inertinite, associated limonitised halos and mottled clay (<5%).

(e) (i) Porphyroskelic grain fabric

(ii) Fine sand grains, well sorted.

(iii) Composition: Oolite : 0 Ooids : 0

Bioclasts : 0 Peloids : 0

Quartz : 5-10 (2-3 shattered quartz)

(f) (i) Pale buff-khaki, with diffuse, warm orange-brown mottles around organics

(ii) Moderately anisotropic microfabric. Bright clay shows 1st order interference greys and yellows.

(iii) Continuous zones surrounding peds. Anisotropic aureoles poorly developed around quartz grains.

(g) Iron-rich mottles surround organics.

(h) Goethite granules in matrix and limonite stains within mottles.

(i) Detrital and ooid quartz grains exhibit cracking and pitting, and some show weathering rinds; stage 1. Shattered quartz are not weathered, stage 0.

HQS12-3

(1) Fairly well-impregnated section, however there has been much surface loss during grinding. Structureless matrix-supported section, with few detrital grains.

(2) (a) Massive to angular-blocky structure

(b) Irregular sinuous pores around peds, large discrete pores within peds.

(c) Rare root-hair passages. These are tiny (0.01mm width, 1mm length) sinuous cross-cutting features, which are coated with limonite-haloes and contain black specks of carbonaceous material in their core.

(d) Unstructured inertinite, with associated limonitised halos (5-7%) and structured cutin.

(e) (i) Porphyroskelic grain fabric

(ii) Fine sand grains, well sorted.

(iii) Composition: Oolite : 0 Ooids : 0
Bioclasts : 0 Peloids : 0
Quartz : 5-7 (1-2 shattered quartz)

(f) (i) Pale buff-brown, khaki in places with warm orange-brown mottles around organics and opaques.

(ii) Moderately anisotropic microfabric. Bright clay shows 1st order interference colours, including greys and yellows around pores and oranges and reds near organics.

(iii) Streaky and fibrous bright clay, surrounding diffuse flecked zones. Anisotropic aureoles poorly developed around quartz and organics.

(g) Iron-rich mottles surround organics.

(h) Goethite granules and limonite stains.

(i) Detrital and ooid quartz grains exhibit pitting and show weathering rinds; stage 1. Shattered quartz are clear, stage 0.

(C) Grey Clay - 'O' horizon

HQS10-4

(1) Poor preparation, weakly impregnated and much loss of section. Only a small fragment of matrix visible, no structures discernible.

(2) (a) Massive structure

(b) Irregular and sinuous pores around peds, large discrete pores within peds.

(c) None recognised - section too poor

(d) Unstructured inertinite, with associated limonitised halos (1-2%)

(e) (i) Agglomeroplasmic grain fabric

(ii) Fine to medium sand grains, moderately sorted.

(iii) Composition: Oolite : 2-5 Ooids : 10-15
Bioclasts : 10-15 Peloids : 5-10
Quartz : 10-15 (2-3 shattered quartz)

(f) (i) Pale brown to speckled reddish brown.

(ii) Weakly anisotropic microfabric. Bright clay shows 1st order interference colours, orange-red around organics and carbonate grains, and greys and yellows around pore edges.

(iii) Diffuse flecked anisotropic zones. Anisotropic aureoles well developed around allochems.

(g) Iron-rich mottles surround organics, nodular and aggregated goethite surrounds some carbonate grains and form impregnations to weathering rinds.

(h) Goethite granules and limonite stains. Equant calcite microspar infilling pore-space.

(i) Allochems show strong weathering rinds with granular inclusions of goethite and ferric hydroxides. The outer lamellae of the ooids are decalcified. Detrital and ooid quartz grains exhibit cracking and pitting, and some show strong weathering rinds developed along edges and into cracks; stage 1. Shattered quartz are not weathered, stage 0.

HQS12-4

(1) Fairly well impregnated sample, some loss of surface during grinding, particularly along pores. Contains abundant organic material and few detrital grains.

(2) (a) Massive in clay-rich areas, to coat and bridge structure in quartz-rich aggregates

(b) Discrete, sinuous pores

(c) None recognised

(d) Rich in unstructured inertinite, with associated limonitised halos (5-10%), also structured cutin and resinite (1-2%).

(e) (i) Agglomeroplasmic (clay-rich) to granular grain fabric (quartz-rich aggregates)

(ii) Fine sand grains, and larger aggregated quartz clusters of fine subangular to subrounded sand within a clay matrix, moderately sorted.

(iii) Composition: Oolite : 0 Ooids : 1-2
Bioclasts : 2-3 Peloids : 0
Quartz : 10-15 (1-2 shattered quartz), up to 50% in aggregates

(f) (i) Speckled buff-brown, darker brown around the abundant opaques

(ii) Weakly anisotropic microfabric. Bright clay shows 1st order interference colours, orange-red around organics and carbonate grains - high iron content

(iii) Striated and fibrous bright clay surround diffuse flecked anisotropic zones. Rare anisotropic aureoles weakly developed around quartz grains. Quartz aggregates show silasepic ('bright sand') fabric (Retallack, 1990).

(g) Iron-rich mottles surround organics, nodular and aggregated goethite surrounds some carbonate grains.

(h) Goethite granules and limonite. Equant calcite microspar nodules infilling pore-space and organics.

(i) Ooids and bioclasts show heavily limonite-stained weathering rinds with granular inclusions and coatings. The outer lamellae of the ooids are decalcified. Detrital, silasepic and ooid quartz grains exhibit cracking and pitting; stage 0(1). Shattered quartz have fresh appearence, stage 0.

(D) Grey Clay - 'B' horizon

HQS9-4

(1) Fairly well-impregnated section, in places there has been loss of matrix during grinding. Matrix-supported, detrital grains present and organics fairly common.

(2) (a) Massive breaking up into angular-blocky structure

(b) Discrete, sinuous and bifurcating pores, and irregular, continuous pores around peds

(c) None recognised

(d) High amount of organics, including unstructured inertinite, with associated limonitised halos (5-10%), including large (2mm length) clast, also structured cutin (?spore) and resinite (1-2%).

(e) (i) Intertextic grain fabric

(ii) Fine to medium sand grains, and granules of oolite, very poorly sorted.

(iii) Composition: Oolite : 5-10 Ooids : 2-5
Bioclasts : 5-7 Peloids : 2-5
Quartz : 20-25 (5-10 shattered quartz)

(f) (i) Speckled buff-brown, darker brown around opaques and organics

(ii) Moderately anisotropic microfabric. Bright clay shows 1st order interference greys and yellows, and oranges and reds near organics.

(iii) Streaky fibrous bright clay. Rare anisotropic aureoles weakly developed around quartz and ooid grains.

(g) Iron-rich mottles surround organics, nodular and aggregated goethite surrounds some carbonate grains and organics.

(h) Goethite granules and limonite stains.

(i) Ooids and bioclasts show weathering rinds with granular limonite/goethite. The outer lamellae of the ooids are decalcified. Oolite fragments exhibit decalcification of ferroan drusy cement to produce limonite and grain-loss porosity. Detrital, ooid and shattered quartz grains exhibit cracking and pitting; stage 0(1).

HQS11-5

(1) Well-impregnated sample. Matrix-supported with abundant detrital grains, including large oolite fragments up to 2-3mm and fine-medium sand.

(2) (a) Massive breaking up into composite laminar-crumb structure in places

(b) Discrete, sinuous and irregular pores.

(c) None recognised.

(d) Rich in unstructured inertinite, with associated limonitised halos (5-10%) which are orientated with the laminations within the soil structure.

(e) (i) Agglomeroplasmic grain fabric.

(ii) Medium-coarse sand, and few large granules of oolite, moderately sorted.

(iii) Composition: Oolite : 5-10 Ooids : 5-10
Bioclasts : 5-7 Peloids : 5-10
Quartz : 10-15 (2-3 shattered quartz)

(f) (i) Buff-brown, orange-brown around organics.

(ii) Moderately anisotropic microfabric. Bright clay shows 1st order interference greys and yellows, orange-red around organics and carbonate grains.

(iii) Striated bright clay zones. Continuous anisotropy around laminar peds. Anisotropic aureoles well developed around carbonate grains and organics.

(g) Iron-rich mottles surround organics, nodular and aggregated goethite surrounds ooids.

(h) Goethite nodules and limonite.

(i) Ooids and bioclasts show weathering rinds with granular limonite/. The outer lamellae of the ooids are decalcified. Oolite fragments exhibit decalcification of ferroan drusy cement to produce limonite and grain-loss porosity. They also have well-developed weathering rinds. Detrital and ooids quartz grains exhibit cracking and pitting; stage 0(1). Shattered quartz are clear, stage 0.

HQS12-5

(1) Fairly well-impregnated sample, with some loss of matrix during grinding. Matrix-supported, with abundant detrital grains and organics. Contaminated heavily with carborundum powder.

(2) (a) Massive structure in matrix-supported areas, coated-grain structure in quartz-rich aggregates.

(b) Rare discrete and irregular pores

(c) None recognised

(d) Rich in unstructured inertinite, with associated limonitised halos (3-5%).

(e) (i) Agglomeroplasmic and intertextic (in quartz-grain aggregates) grain fabrics

(ii) Fine-medium sand, and few large granules of oolite, poorly sorted.

(iii) Composition: Oolite : 2-3 Ooids : 1-2
Bioclasts : 2-3 Peloids : 2-3
Quartz : 10-15 (shattered quartz 2-5)

(f) (i) Buff-brown, darker brown around organics

(ii) Weakly anisotropic microfabric. Bright clay shows 1st order interference greys and yellows, and orange-red around organics and carbonate grains.

(iii) Striated and fibrous anisotropic zones surround zones of diffuse flecked bright clay. Rare anisotropic aureoles weakly developed around quartz grains. Quartz aggregates show bright sand anisotropy.

(g) Iron-rich mottles surround organics, nodular and aggregated goethite surrounds carbonate grains and impregnate weathering rinds.

(h) Goethite granules and limonite stains; rare equant calcite infills pores and organics.

(i) Ooids and bioclasts show weathering rinds with granular limonite/. The outer lamellae of the ooids are decalcified. Oolite fragments exhibit decalcification of ferroan drusy cement to produce limonite and grain-loss porosity. They also have well-developed weathering rinds. Detrital and ooid quartz grains exhibit cracking and pitting; stage 0(1). Shattered quartz are fresh, stage 0.

(E) Grey Clay - 'BC' and 'C' horizons

HQS12-6

(1) Well impregnated section, rare grain-loss during preparation. Quite granular with much detrital grains and organics present.

(2) (a) Massive, in places angular-blocky structure

(b) Elongate, sinuous and bifurcating pores define blocky peds

(c) None recognised

(d) Rich in unstructured inertinite, with associated limonitised halos (10-15%). These occur orientated parallel to the elongate pores or 'bedding planes' within the soil section. Resinite and cutin also present.

(e) (i) Granular grain fabric toward intertextic in places

(ii) Fine-coarse sand, moderately sorted.

(iii) Composition: Oolite : 10-15 Ooids : 5-7

Bioclasts : 10-15 Peloids : 5-10

Quartz : 15-20 (1-2 shattered quartz)

(f) (i) Buff-brown and khaki-brown with darker orange-brown mottles around organics.

(ii) Weakly anisotropic microfabric. Bright clay shows 1st order interference greys and yellows with orange-red around organics and carbonate grains.

(iii) Striated bright clay and regions of flecked anisotropy. Skelsepic aureoles weakly developed.

(g) Iron-rich mottles surround organics, nodular and aggregated goethite surrounds ooids.

(h) Goethite granules and limonite. Void-filling calcite drusy spar within organic moulds.

(i) Carbonate grains show weathering rinds with granular limonite/goethite. The outer lamellae of the ooids are decalcified. Oolite fragments exhibit decalcification of ferroan drusy cement to produce limonite and grain-loss porosity. Quartz grains exhibit and pitting; stage 0.

APPENDIX C6 ; Description of petrographic and polished sections, and acetate peels of the Chipping Norton Formation limestones altered by karstification

Each specimen was examined in three ways to determine the rock type and degree of alteration to the original limestone during karst formation:

(1) Hand specimens and polished blocks were examined in conjunction with stained acetate peels to facilitate recognition of the large scale textural features and

general mineralogy of the rock type. Features looked for and recognised at this point are similar to those used for the description of unaltered limestone and include:

(a) Textural features such as any inherited lamination or bedding from the parental limestone, and any new solutional features such as fissures and veining.

(b) Overall grainsize distribution and inclusion of any large distinct clasts or lenses of clasts (see above).

(c) Colour variation and weathering. Colour is described and also given a reference Munsell^{R} value (see above). The limestones taken from the karst region often show a degree of colour mottling and marbling. In the following descriptions different coloured mottles are described separately.

(d) Approximate determination of the amount of cementation within the rock. This could either be primary or karstic cementation.

(2) Examination of thin section in natural daylight, in conjunction with acetate peels. This helps to recognise any distinct colour mottling or banding and areas of distinct weathering or alteration within the thin section before description is attempted. Also recognised at this point are the general grainsize distribution and sorting within the thin section and the inclusion of distinguishable zones of cementation.

(3) The thin section and acetate peels are then described fully under the petrographic microscope at x40, x100 and x400 magnification, in both plane and cross-polarised transmitted light. In most thin sections there are two or more distinct regions or mottles, the first being areas of unaltered limestone (section 4.4.1). These are described and classified in the standard manner (see above). Other areas which are usually heavily stained yellow-brown or dark brown are described separately. These are the regions of the limestone which have undergone alteration during the period of karstification, they exhibit a mixture of features inherited from the parent rock, such as composition of grains and cement, and others formed during this period, such as weathering products and new cements. The karstified limestones can be considered to be the regolith to the overlying paleosols which have formed from more advanced subaerial alteration of the limestones. Hence in attempting a simple classification scheme of these limestones, I have utilised the most applicable characters from both the limestone and soil descriptions set out in the preceding appendices (see above); these are:

(a) Inherited texture and composition. Based upon the Folk and Dunham schemes for limestones and the Retallack scheme for soils outlined above, this aims to illuminate any difference between the altered and parental limestones. This includes: (i) Composition and proportions of detrital grains, cement and matrix (in modal percentages)

(ii) Texture and distribution of grains.

(iii) Grainsize and sorting.

(b) Original preservation of grains, including:

(i) Micritic envelopes and borings around the edges of allochems such as bioclasts and ooids.

(ii) Calcitisation or *in situ* replacement of aragonitic allochems. Any preserved compositional zoning of ferroan-rich ooids or internal texture of bioclasts.

(iii) Dissolution and replacement by ferroan drusy sparite.

(c) Degree of alteration of grains

(i) Alteration products of carbonate grains by dissolution such as residues of ferric hydroxide and aggregated iron oxides.

(ii) Degree of alteration and weathering based upon the scheme devised in FitzPatrick (1984, 1994) (see above).

(d) Cementation history - inherited. The four types of carbonate cement seen within the unaltered limestones are also found in the karst specimens, they are

(i) Pore-filling micrite or calcite microspar.

(ii) Radial-fibrous pore-lining calcite and syntaxial overgrowths of high magnesium calcite upon allochems

(iii) Drusy mosaic pore-filling ferroan calcite spar.

(iv) Equant calcite vein-spar.

(e) Karstic alteration of primary cements

(i) Oxidation products and secondary minerals, for example granules of pore-filling goethite and limonite formed from the oxidation during weathering of ferroan calcite cements. Manganese dioxide or pyrolusite is also found.

(ii) Matrix and grain coatings; this includes a description of the colour in plane-polarised light, composition and microfabric or texture of the matrix based upon the scheme employed for soil classification (see above). The matrix and coatings are usually composed of ferric oxide/hydroxide.

(f) Cementation history - post-karst cements.

(g) Porosity and structure. Porosity can be a primary depositional feature of the limestone (see above) or may have formed from secondary dissolution as a result of diagenesis and sub-aerial karstic weathering. The structure refers to the total organisation and spatial distribution of the thin section, in terms of soil micromorphology

(A) Altered limestones from around the palaeokarstic surface at Hornsleasow

HQCN2 - taken from base of the clay-filled hollow, 1990 by author

(1) Fairly well-cemented (except for top 3-7mm) calcarenite. Shows distinct colour zonations or mottling, particularly towards the top of the specimen. Mottles are pale-creamy white (10YR 8/2) well-cemented with ferroan calcite (mauve) irregular shaped areas, measuring 10-50mm. These are surrounded by either thin (1-2mm), decalcified and friable dark brown (10YR 4/2) areas, associated with much iron oxide and no staining, or diffuse browny-orange (10YR 6/6) areas, which show pink stained (calcitic) friable cemented and ferric oxyhydrite-rich decalcified regions.

(2) Two thin sections prepared.

HQCN2-1 - diffuse marbled zone 50mm below surface

Distinct (mottling) and diffuse (marbled) appearence of colours correspond with those seen in polished block. Brown areas are cemented ironstones, showing some calcitic cement (pink stained), pale areas are remnant limestone with ferroan calcite cement (mauve stained).

HQCN2-2 - Near surface, decalcified and mottled zone

Very distinct mottled appearence, showing small (2-5mm) orange brown areas surrounded by dark-brown boundaries. Mottling appears to be following bedding or laminations. Toward the top of the specimen becomes more marbled with pale cream areas grading over about 10mm into diffuse orange-brown areas. Much decalcification within the dark-brown ironstone areas. The remnant limestone areas show ferroan calcite cement.

(3) Thin section description.

HQCN2-1 - pale cream mottles

(a) Grainstone

(b) Modal composition (percentages) : oobiosparite

Matrix (carbonate mud) : 0-5

Cement (sparite) : 45-50

Allochems : 45-55	Ooids : 20-25	Bioclasts : 15-20
	Peloids : 10-15	Intraclasts : 0
Other grains : 5-10	Quartz : (2-5 ooids)	Opaques : 5-7

(c) Clasts rounded, well sorted - rounded oobiosparite

(d) (ii) <10% of pore spaces show radial fibrous calcite cement and rare syntaxial overgrowths upon echinoid clasts.

(iii) Pore-filling drusy ferroan calcite

(iv) Rare vein-filling equant calcite

(e) Low interparticle and intercrystal porosity - 5% clasts lost during preparation

(f) (i) Micritic envelopes on >95% allochems and 65-70% allochems are partially or wholly micritised.

(ii) Calcitisation of ooids. Around 1-2% of ooids show well-preserved concentric and radial structures. These are compositionally zoned from an inner region of micritic lamellae to outer lamellae composed of ferroan micrite and siderite.

(iii) Wholesale dissolution of bioclasts and incomplete replacement by ferroan drusy spar (30-35%).

(iv) Incomplete dissolution of carbonate cements and especially pore-filling ferroan calcite to leave a residue of ferric hydroxide and aggregated iron oxides and strong weathering rinds. Micritised bioclasts show dissolution at their edges and ooids contain opaque inclusions in their outer lamellae.

HQCN2-1 - Pale brown boundaries

(a) (i) Modal composition (percentages) : altered ferruginous oobiosparite

Matrix (carbonate mud) : 10-15

Cement (sparite) : 5-10

Allochems :	Ooids : 10-15	Bioclasts : 10-15
	Peloids : 5-10	Intraclasts : 5-10
Other grains :	Quartz : 10-15	Opaques : 20-25

(ii) Texture and distribution : grainstone, intertextic distribution

(iii) Medium to coarse sand, moderately well sorted

(b) (i) Micritic envelopes are seen on 2-5% of well-preserved bioclasts. Most allochems are micritised, visible texture seen in about 25-30%.

(iii) Ferroan drusy sparite is preserved in 5-10% bioclasts.

(c) (i) Ferroan drusy calcite infilled bioclasts are dissolved leaving central voids and inclusions of limonite and goethite. Ooids and peloids show inclusions of goethite and limonite, these are particularly concentrated in the outer lamellae of ooids, indicating that these were once iron-rich calcite.

(ii) Ooids and bioclasts have well-developed weathering rinds. Calcitised bioclasts are better preserved but show alteration along cracks and thin weathering rinds. Ferroan drusy-infilled bioclasts have developed around the edges thick micritised weathering rinds. The higher proportion of non-ooid quartz grains, suggests that there has been complete dissolution of some superficial ooids. These quartz grains show pitting and cracking; stage 0(1).

(d) (iii) Rare drusy mosaic pore-filling ferroan calcite spar.

(iv) Equant calcite vein-spar.

(e) (i) Drusy ferroan-calcite has been largely oxidised to leave residual ferric oxides/hydroxides within the voids. In other places cement remains but

contains inclusions of granular goethite and limonite. Gives overall yellow appearence. No early cements such as syntaxial overgrowths or radial-fibrous calcite remain intact.

(ii) Opaque coatings of goethite and pyrolusite, and reddish-opaque granules of hematite and goethite set within a yellow-brown fine-grained matrix of limonite (opaque) and micrite.

(iii) Large equant calcite cement may be case-hardening cement.

(f) No post-karst cementation

(g) High interparticle and intercrystal porosity - 10-15% clasts lost during preparation in areas not cemented by late calcite cement. Pores are discrete and irregular. Composite single-grain, and coat and bridge structure

HQCN2-2 - pale cream mottles

(a) Grainstone

(b) Modal composition (percentages) : oobiosparite

Matrix (carbonate mud) : 3-5

Cement (sparite) : 40-45

Allochems : 45-50	Ooids : 20-25	Bioclasts : 15-20
	Peloids : 10-15	Intraclasts : 2-5
Other grains : 1-2	Quartz : (2-5 ooids)	Opaques : 1-2

(inclusions up to 10%)

(c) Clasts rounded, well sorted - rounded oobiosparite

(d) (ii) <5% of pore spaces show radial fibrous calcite cement; rare syntaxial overgrowths on echinoid fragments.

(iii) Pore-filling drusy ferroan calcite common

(iv) Large cross-cutting veins of calcite which show evidence of pressure solution of allochems; do not cut though opaque-rich yellow-brown mottled areas, therefore they are pre-karstic solutional features or remains of a case-hardening cement.

(e) Low interparticle and intraparticle porosity - 5-10% clasts lost during preparation

(f) (i) Micritic envelopes on 65-70% allochems. 55-60% of allochems are micritised.

(ii) Calcitisation of ooids and 2-5% bioclasts.

(iii) Rare wholesale dissolution of bioclasts and incomplete replacement by ferroan drusy spar (15-20%).

(iv) Incomplete dissolution of carbonate cements and especially pore-filling ferroan calcite to leave a residue of ferric hydroxide and aggregated iron oxides and strong weathering rinds. Calcitised bioclasts show

dissolution at their edges. Many opaques associated with compositionally-zoned ooids.

HQCN2-2 - brown-yellow mottles

(a) (i) Modal composition (percentages) : altered ferruginous biopelsparite

Matrix (carbonate mud) : 5-10

Cement (sparite) : 2-5

Allochems : 45-50	Ooids : 5-10	Bioclasts : 15-20
	Peloids : 15-20	Intraclasts : 5-10
Other grains :	Quartz : 5-7	Opaques : 25-30 (matrix)

(ii) Texture and distribution : grainstone, granular-intertextic distribution. Point-to-point contact with softer grains pushed into each other and some pressure solution along contact.

(iii) Medium sand, well sorted

(b) (i) Micritic envelopes seen on 15-20% bioclasts only, up to 65% are partially or wholly micritised.

(ii) Calcitisation of ooids.

(iii) Ferroan drusy sparite seen in only 5-10% of bioclasts

(c) (i) Ferroan drusy calcite infilled bioclasts are partially dissolved leaving voids and ferric oxyhydrite grains. Calcitised allochems show inclusions of goethite and limonite, which are particularly concentrated in the outer lamellae of ooids, indicating that these were once iron-rich calcite or siderite.

(ii) Ooids have well-developed weathering rinds composed of dirty-brown micrite and in some places are completely unrecognisable. Calcitised bioclasts are better preserved than drusy-replaced ones, which have largely dissolved to an opaque residue, but show alteration along cracks. The higher proportion of non-ooid quartz grains, suggests that there has been complete dissolution of some superficial ooids. These quartz grains show pitting and cracking, and opaque-bearing weathering rinds; stage 0(1).

(d) (iii) Drusy mosaic pore-filling ferroan calcite spar, <5%.

(e) (i) Drusy ferroan-calcite has been largely oxidised to leave residual ferric oxyhydrites within the voids. Gives overall yellow appearence.

(ii) Opaque coatings of goethite (reddish) and pyrolusite (black), and reddish-opaque granules of hematite and goethite set within a reddish-brown fine-grained matrix of limonite and carbonate mud (from partially dissolved peloids).

(f) None-recognisable.

(g) High interparticle and intercrystal porosity - 20-30% clasts lost during preparation in areas not cemented by late calcite cement. Pores are discrete and irregular. Coat and bridge structure.

HQCN2-2 Dark-brown vein-like areas (basal region)

(a) (i) Modal composition (percentages) : calcareous ironstone

Matrix (carbonate mud) : 1-2

Cement (sparite) : 2-5

Allochems : 15-20 Ooids : 1-2 Bioclasts : 10-15

Peloids : 5-10 Intraclasts : 0

Other grains : Quartz : 10-15 Opaques : 40-45 (matrix)

(ii) Texture and distribution : grainstone-floatstone, intertextic to agglomeroplasmic distribution

(iii) Fine-medium sand, moderately well sorted

(b) (ii) Micritisation of allochems around 15-20%

(c) (i) Ferroan drusy calcite infilled bioclasts are completely dissolved leaving voids and ferric oxyhydrite grains. Micritised and low magnesium calcite bioclasts show inclusions and coatings of goethite and limonite. Opaques are concentrated in the lamellae of ooids and the interior of peloids indicating that these were once iron-rich calcite.

(ii) Allochems have well-developed weathering rinds. Some low magnesium calcite bioclasts are better preserved than other allochems but show alteration along cracks. The higher proportion of non-ooid quartz grains, suggests that there has been complete dissolution of many superficial ooids. These quartz grains show pitting and cracking, and large weathering rinds; stage 1.

(d) (iii) Drusy mosaic pore-filling ferroan calcite spar is extremely rare.

(e) (i) Drusy ferroan-calcite has been almost completely oxidised to leave residual ferric oxides/hydroxides within the voids. Gives overall dark brown appearence.

(ii) Opaque coatings of goethite and pyrolusite, and reddish-opaque granules of hematite and goethite set within a dark brown fine-grained matrix of limonite. Forms an ironstone.

(f) None seen

(g) High interparticle and intercrystal porosity - 10-15% clasts and cements lost during preparation in areas not cemented by late calcite cement. Pores are numerous discrete and irregular. Composite massive and coat and bridge structure

HQCN3 - taken from karstic surface below the unexcavated clay unit, 1989 by M.J. Simms.

(1) Moderate-cemented calcarenite. Top 30-50mm of block shows much dissolution of grains and cement and a diffuse region of colour marbling. The rest of the specimen has distinct colour zonations or mottling, particularly towards the base of the specimen. Mottles are pale-creamy white (10YR 8/2) well-cemented with ferroan calcite (mauve) irregular shaped areas, measuring 10-50mm. These are surrounded by either thin (1-2mm), decalcified dark brown ironstone (10YR 4/2) areas, associated with much iron oxide and no staining, or diffuse friable browny-orange (10YR 6/6) areas, which show pink stained (calcitic) friable cemented and ferric hydroxide-rich decalcified regions.

(2) Two thin sections prepared.

HQCN3-1 - basal region , some 100mm below surface

Well-cemented clearly mottled region. Two different coloured zones clearly delineated, large (>10mm) rectangular pale yellowish areas of equigranular calcarenite, surrounded by well-defined dark brown and reddish-brown vein-like borders of ironstone (1-2mm in width).

HQCN3-2 - taken from 20-30mm below surface of block, in the marbled decalcified region

Poorly-cemented limestone, which 20-30% grain loss. Shows marbling with two colour bands delineated, the dark-brown diffuse region of ironstone (2-3mm width) with significant porosity and loss of cement and grains. The second is a pale-brown to orange region which contains many opaques, but is compositionally an equigranular calcarenite.

(3) HQCN3-1- pale cream-yellow mottles

(a) Grainstone, in places a rudstone

(b) Modal composition (percentages) : oobiosparite

Matrix (carbonate mud) : 2-5

Cement (sparite) : 40-45

Allochems : 35-40 Ooids : 20-25 (rudstone 5-10) Bioclasts : 15-20 (rudstone 25-30)

Peloids : 10-15 Intraclasts : 2-5

Other grains : 10 Quartz : (2-5 ooids) Opaques : 2-5 (rudstone 5-10)

(c) Clasts rounded, well sorted - rounded oobiosparite

(d) (ii) <5% of pore spaces show radial-fibrous calcite cement; rare syntaxial overgrowths

(iii) Pore-filling drusy ferroan calcite

(e) Moderate interparticle, intercrystal and mouldic porosity - 15-20% clasts lost during preparation, especially high in grainstone.

(f) (i) Micritic envelopes on 55-60% allochems and 60-65% allochems exhibit partial or complete micritisation.

(ii) Calcitisation of ooids. Around 2-5% of ooids show well-preserved radial and concentric structures and are compositionally zoned from an inner micritic rich zone to lamellae composed of ferroan micrite and siderite.

(iii) Rare wholesale dissolution of bioclasts and incomplete replacement by ferroan drusy spar (15-20%).

(iv) Much opaques associated with compositionally zoned ooids. Some of these grains show obvious signs of dissolution and alteration at their edges. Many have developed weathering rinds which contain abundant ferric oxyhydrite opaques.

CNHQ3-1 - dark-brown borders

(a) (i) Modal composition (percentages) : calcareous ironstone

Matrix (carbonate mud) : 2-5

Cement (sparite) : 2-5

Allochems : 20-25	Ooids : 5-10	Bioclasts : 10-15
	Peloids : 5-10	Intraclasts : 1-2
Other grains :	Quartz : 2-5	Opaques : 45-50

(ii) Texture and distribution : agglomeroplasmic distribution

(iii) Medium to coarse sand, moderately well sorted

(b) (i) Micritic envelopes upon only 1-2% and 45-50% allochems are partially or wholly micritised.

(ii) Calcitisation of all ooids.

(c) (i) Ooids and peloids show inclusions and coatings of goethite and limonite, these are particularly concentrated in the outer lamellae of ooids and also occur in the weathering rinds of micritised bioclasts.

(ii) Allochems have well-developed weathering rinds. Low magnesium calcite bioclasts are better preserved than drusy-replaced ones which have been totally dissolved away. Those that occur within the limestone next to the dark-brown areas have been completely destroyed within these solutional boundaries. The higher proportion of non-ooid quartz grains, suggests that there has been complete dissolution of some superficial ooids. These quartz grains show pitting and cracking, and some weathering rinds; stage 0(1).

(d) (iii) Drusy mosaic pore-filling ferroan calcite spar.

(e) (i) Drusy ferroan-calcite has been almost completely oxidised to leave residual ferric oxides/hydroxides as the matrix. Gives overall brown-opaque appearence.

(ii) Matrix consists of a dark-brown opaque mineral (limonite) with black (?pyrolusite) and reddish-brown (goethite/hematite) inclusions.

(f) None seen

(g) High interparticle, intercrystal and vug porosity - 10-15% clasts lost during preparation in areas not cemented by late calcite cement. Pores are frequent, discrete and irregular. Coat and bridge structure, almost massive.

HQCN3-2 - Pale yellow-brown marbled region

(a) Grainstone

(b) Modal composition (percentages) : oobiosparite

Matrix (carbonate mud) : 2-5

Cement (sparite) : 45-50

Allochems : 40-45	Ooids : 15-20	Bioclasts : 15-20
	Peloids : 10-15	Intraclasts : 2-5
Other grains : 10	Quartz : (2-3 ooids)	Opaques : 5-10

(c) Clasts rounded, well sorted - rounded oobiosparite

(d) (iii) Pore-filling drusy ferroan calcite

(e) Low interparticle and intercrystal porosity - 5-10% clasts lost during preparation

(f) (i) Micritic envelopes on 75-80% allochems. 55-60% of grains are wholly or partially micritised.

(ii) Calcitisation of ooids. Around 5-10% of ooids are compositionally zoned from inner micritic lamellae to outer lamellae composed of ferroan micrite and siderite. These show well-preserved radial and concentric structures

(iii) Rare wholesale dissolution of bioclasts and incomplete replacement by ferroan drusy spar (5-10%).

(iv) Much opaques associated with compositionally zoned ooids occur as weathering rinds around the outer lamellae. Also granular ferric oxyhydrites form coatings to bioclasts which have strong weathering rinds. Some of these grains show obvious signs of dissolution and alteration at their edges and contain inclusions of granular goethite.

HQCN3-2 - dark-brown decalcified region

(a) (i) Modal composition (percentages) : ferruginous oobiosparite

Matrix (carbonate mud) : 2-5

Cement (sparite) : 2-3

Allochems : 30-35	Ooids : 10-15	Bioclasts : 10-15
	Peloids : 5-10	Intraclasts : 2-5
Other grains :	Quartz : 5-10	Opaques : 35-40

(matrix)

(ii) Texture and distribution : grainstone, granular-intertextic distribution

(iii) Medium sand, well sorted

(b) (i) Micritic envelopes seen on around 5-10% bioclasts. 50-55% of allochems are micritised.

(ii) Calcitisation of all ooids

(c) (i) Ferroan drusy calcite infilled bioclasts are dissolved leaving voids and limonitised matrix. Ooids show inclusions of goethite and limonite, these are particularly concentrated in the outer lamellae, indicating that these were once iron-rich calcite and siderite. Peloids and some bioclasts contain opaque inclusions and all allochems are coated.

(ii) Allochems and quartz grains have well-developed reddish-brown and brown weathering rinds, containing many opaques including red granules of hematite. Low magnesium calcite bioclasts are better preserved, but show alteration along cracks and strong weathering rinds. Soft grains show grain-to-grain contact solution. The higher proportion of non-ooid quartz grains, suggests that there has been complete dissolution of some superficial ooids. These quartz grains show pitting and cracking; stage 0.

(d) (iii) Remnant drusy mosaic pore-filling ferroan calcite spar.

(e) (i) Drusy ferroan-calcite has been largely oxidised to leave an opaque matrix of iron hydroxides and oxides within the voids. In other places cement remains but contains inclusions of granular goethite and limonite.

(ii) Opaque coatings of reddish-brown coalesced granules of limonite and goethite. Dark brownish-red and brownish-yellow opaque matrix of limonite and goethite, with red granules of hematite.

(f) None seen

(g) High interparticle and intercrystal - 70-80% clasts lost during preparation in areas cemented by ferric oxyhydrites. Pores are discrete and irregular. Coat and bridge structure.

HQCN4 - karstic limestone taken from side wall of hollow

(1) Well-cemented sparry calcarenite. Pale yellowish (10YR 8/6) in colour, appears largely unweathered and unaltered.

(2) Equigranular well-cemented limestone. No colour banding and appears unaltered by karstic processes. Disseminated brown-yellow opaques visible.

(3) (a) Grainstone

(b) Modal composition (percentages) : oobiosparite

Matrix (carbonate mud) : 5-10

Cement (sparite) : 45-50

Allochems : 35-40 Ooids : 15-20 Bioclasts : 15-20

Peloids : 10-15 Intraclasts : 3-5

Other grains : <5 Quartz : (2-5 ooids) Opaques : 2

(c) Clasts rounded, moderate sorted - rounded oobiosparite

(d) (ii) Remnant fibrous-radial calcite (<5%) pore-lining.

(iii) Pore-filling drusy ferroan calcite

(iv) Equant non-ferroan calcite (pink stained) infilling large vugs and veins, may be case-hardened cement that pervades the karstic surface, contains no opaques.

(e) Very low intercrystal and intraparticle porosity - 1-2% clasts lost during preparation

(f) (i) Micritic envelopes clearly visible on 85-90% allochems and 45-50% allochems wholly or partially micritised.

(ii) Calcitisation of ooids. Ooids are well-preserved and show excellent concentric zoning and radial structure. These are also compositionally zoned from inner lamellae composed of micrite to outer lamellae composed of ferroan micrite and siderite.

(iii) Rare wholesale dissolution of bioclasts and incomplete replacement by ferroan drusy spar (5-10%).

(iv) Some allochems show weathering rinds containing brown and reddish-brown opaques. Some of these grains show obvious signs of dissolution and have developed distinct weathering rinds of alteration at their edges which contain inclusions of ferric hydroxides and oxides. There are abundant tiny microspherules of red-brown goethite/hematite within the central regions of peloids, the outer lamellae of ooids and within weathering rinds developed upon bioclasts.

(B) Limestone clasts found within the Grey Clay directly overlying the palaeokarstic surface

HQP1-karstic pebble picked up from the floor of the hollow by M.J. Simms, in 1989

(1) 3x5cm sub-elliptical pebble, with sub-rounded appearence. Outer edges heavily stained brown-yellow (10YR 6/6) which pervades into the interior by about 1-5mm. Interior shows an equigranular calcarenite. Well-cemented dark grey (5YR 6/1) non-corroded interior.

(2) Equigranular sparry-cemented grey-brown coloured specimen. Much dark brown staining at edges and increased amount of opaques.

(3)

HQP1 - interior of pebble

(a) Wackestone

(b) Modal composition (percentages) : oopelsparite

Matrix (carbonate mud) : 20-25

Cement (sparite) : 20-25

Allochems : 35-40 Ooids : 15-20 Bioclasts : 5-10

Peloids : 10-15 Intraclasts : 0

Other grains : 10-15 Quartz :3-5 Opaques : 5-10

(c) Clasts rounded, well sorted - rounded poorly washed oopelsparite

(d) (i) Micrite may well be a secondary cement rather than original carbonate mud

(iii) Pore-filling drusy ferroan calcite

(iv) Vug-filling drusy non-ferroan calcite - possibly case-hardening cement

(e) Low interparticle and intercrystal porosity - 5-10% clasts lost during preparation

(f) (i) Micritic envelopes on 65-70% allochems and 40-45% allochems are micritised.

(ii) Calcitisation of ooids. Ooids are compositionally zoned from inner micritic lamellae to outer lamellae composed of ferroan micrite and siderite.

(iii) Rare wholesale dissolution of bioclasts and incomplete replacement by ferroan drusy spar (5-10%).

(iv) Much opaques associated with compositionally zoned ooids occur as weathering rinds around the outer lamellae. Also ferric oxyhydrites form coatings to bioclasts and lines of inclusions within grains. Drusy-filled bioclasts show obvious brown weathering rinds and coatings, which contain inclusions of ferric hydroxides and oxides.

<u>HQP1 - case-hardened weathering-rind</u>

(a) (i) Modal composition (percentages) : ferruginous oopelsparite

Matrix (carbonate mud) : 2-5

Cement (sparite) : 15-20

Allochems : 10-15 Ooids : 2-5 Bioclasts : 5-10

Peloids : 2-5 Intraclasts : 0

Other grains : Quartz : 10-15 Opaques : 40-45 (matrix)

(ii) Texture and distribution : grainstone, intertextic distribution

(iii) Medium to coarse sand, moderately well sorted

(b) (i) Micritisation of up to 70% of allochems

(ii) Ooids are calcitised.

(iii) Ferroan drusy sparite seen in about 5% allochems.

(c) (i) Ferroan drusy calcite infilled bioclasts are partially or wholly dissolved leaving limonite matrix and goethite grains. Ooids show inclusions of goethite and limonite, these are particularly concentrated in the outer lamellae, indicating that these were once iron-rich calcite. Bioclasts and peloids show lines of goethite inclusions and coatings of limonite.

(ii) Allochems have well-developed weathering rinds. Low magnesium calcite bioclasts are better preserved than drusy-replaced ones, which have largely dissolved to an opaque residue, but show alteration along cracks. The higher proportion of non-ooid quartz grains, suggests that there has been complete dissolution of some superficial ooids. These quartz grains show pitting and cracking; stage 0(1).

(d) (iii) Drusy mosaic pore-filling ferroan calcite spar.

(iv) Equant calcite vein-spar.

(e) (i) Drusy ferroan-calcite has been largely oxidised to leave residual ferric oxyhydrites within the voids. In other places cement remains but contains inclusions of granular goethite and limonite. No early cements such as syntaxial overgrowths or radial-fibrous calcite remain intact.

(ii) Opaque coatings of goethite and pyrolusite, and reddish-opaque granules of hematite and goethite set within a dark-brown fine-grained limonitic matrix. Gives overall dark yellow appearence of matrix

(f) None seen

(g) High interparticle and intercrystal porosity - 5-10% clasts lost during preparation. Pores are discrete and irregular. Composite single-grain, and coat and bridge structure.

<u>HQP2- limestone clast found on top of the karstic surface by M.J. Simms, 1989 (post-excavation).</u>

(1) Rubbly, friable small clast (30x40mm) of matrix-supported calcilutite. Pale buff (10YR 8/2) weathering to a warm pale orange (10YR 8/6). Unlike any of the surrounding karstic Chipping Norton limestones and appears to be largely unaltered by dissolution or case-hardening upon the karst and by chemical weathering within the soil. May be derived during excavation or from cambering cracks from overlying Chipping Norton oomicrites or within Grey Clay soil formation.

(2) Brown fine-grained muddy matrix with very few grains. Contains large pores, which appear to be birdseyes.

(3) (a) Carbonate mudstone

(b) Modal composition (percentages) : ooid-bearing fossiliferous dismicrite

Matrix (carbonate mud) : 75-85

Cement (sparite) : 5-10 (in fenestrae)

Allochems : 5-10	Ooids : 2-5	Bioclasts : 2-3
	Peloids : 1-2	Intraclasts : 0
Other grains : 3-5	Quartz : 1-2	Opaques : 2-5

(c) Clasts rounded, well sorted - rounded ooid-bearing fossiliferous dismicrite

(d) (iii) Pore-filling drusy ferroan calcite

(e) Fenestral porosity - 5-10% clasts lost during preparation

(f) (i) Micritic envelopes on 55-60% allochems.

(ii) Calcitisation of ooids. Compositionally zoned ooids show range from inner micritic lamellae to outer ferroan micrite and siderite bearing lamellae.

(iv) Much opaques associated with compositionally zoned ooids occur within weathering rinds and as coatings around the outer lamellae. Also ferric oxyhydrites form coatings to bioclasts.

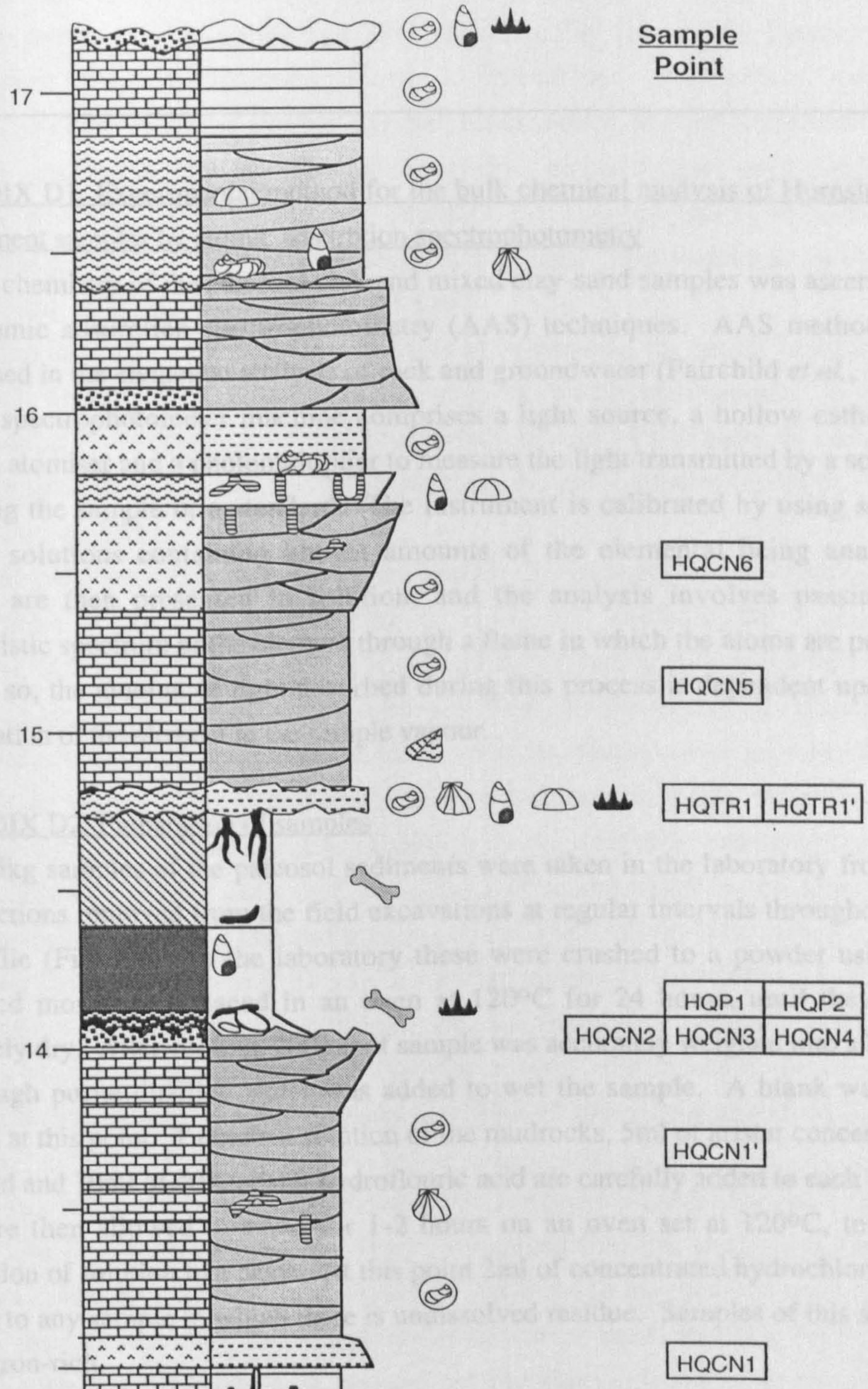


Figure C1. Graphic log of the Hornsleasow section showing position of sampling points

Appendix D

S.J. Metcalf, 1995

APPENDIX D1: Experimental method for the bulk chemical analysis of Hornsleasow lens sediment samples by atomic adsorption spectrophotometry

The bulk chemistry of the paleosol clay and mixed clay-sand samples was ascertained using atomic adsorption spectrophotometry (AAS) techniques. AAS methods are widely used in the elemental analysis of rock and groundwater (Fairchild *et al.*, 1990). The AA spectrophotometer machine comprises a light source, a hollow cathode, a flame, an atomiser and a photomultiplier to measure the light transmitted by a solution containing the sample or a standard. The instrument is calibrated by using several standard solutions containing known amounts of the elemental being analysed. Samples are then presented in solution, and the analysis involves passing the characteristic spectrum of the element through a flame in which the atoms are present. In doing so, the amount of light adsorbed during this process is dependent upon the concentration of the element in the sample vapour.

APPENDIX D2: Preparation of samples

Fresh 0.5kg samples of the paleosol sediments were taken in the laboratory from the block-sections removed from the field excavations at regular intervals throughout the soil profile (Fig. C2). In the laboratory these were crushed to a powder using an automated mortar and placed in an oven at 120°C for 24 hours, until they were completely dry. After cooling, 200mg of sample was accurately weighed into a beaker and enough purite distilled water was added to wet the sample. A blank was also prepared at this point. T obtain a solution of the mudrocks, 5ml of aristar concentrated nitric acid and 10ml of 40% aristar hydroflouric acid are carefully added to each vessel. These are then allowed to stand for 1-2 hours on an oven set at 120°C, to allow evaporation of solution to a paste. At this point 2ml of concentrated hydrochloric acid is added to any sample in which there is undissolved residue. Samples of this sort are usually iron-rich.

After cooling, 5ml concentrated nitric acid and enough water to facilitate complete dissolution of the paste is added to the vessel. This solution is then left to stand on a warm oven for 10 minutes for dissolution to take place. After cooling, the sample solution is placed within a 100ml flask and made up to 100ml with purite de-ionised distilled water.

APPENDIX D3: Operation and calibration of instrument

After initial start up of the AAS machine and fume-extractor hood, the correct lamp is selected by pressing the button marked "GOTO" (2, on Fig. D1) and the number of the lamp required, this will display the element to be analysed. All element lamps are preset with the correct atomic number and maximum current, and are focused to give a maximum negative deflection. The air compressor and acetylene gas cylinder are turned on, with the gas pressure set to 9lb/in2. For the analysis of aluminium, a nitrous oxide gas cylinder was also employed at this stage. The instrument should then be left to warm up for approximately 20 minutes.

After this time has elapsed, the flame is lit by pressing the 'ignite' button on the machine (5, on Fig. D1). Purite distilled water is run through machine to enable stabilisation and equilibriation of the flame, which should turn to blue. In case of an emergency the flame can be turned off by pressing the 'off' button in the flame control panel (5, on Fig. D1).

The machine is then ready to be calibrated. The number of standards used (in this case three) is keyed into the machine (6, on Fig. D1), including the top standard and loaded. After this the top standard concentration (7, on Fig. D1) is loaded into the machine (in weight percentages). The following steps involve keying into the machine the concentrations of the remaining standard solutions, this involves pressing the 'set concentration' button (8 and 9, on Fig. D1) before each entry is made. After this stage the 'X' button on the standard control panel (10, on Fig. D1) will repeatably flash and the top standard solution must then be introduced to the machine, by placing the solution under the flame nozzle and pressing 'RUN' (12, on Fig. D1). When this run is complete the '0' button will flash on the standard control button and in this case the blank sample will be introduced to the machine. This procedure continues until all the standards have been run through the machine in turn. Purite water should be run through the machine after each solution to eliminate cross-contamination between standards.

After this procedure the instrument will construct a calibration curve for the element and the 'RUN' button will flash, indicating that the analysis is ready to commence. At this point the blank should be run through the machine once more to check for instrumnetal drift, if at this point the instrument needs resetting the 'BLANK' button (13, on Fig. D1) should be pressed and the blank introduced to the machine again. This step will re-zero the calibration. Ideally, the blank solution should be run through the machine after every couple of samples and the zero re-set as necessary. Each sample should then be introduced to the machine in turn and analysed at least in triplicate. The results (Appendix D5) are the mean of these analyses. However, the hazard involved with running the combined nitrous oxide-acetylene flame for the

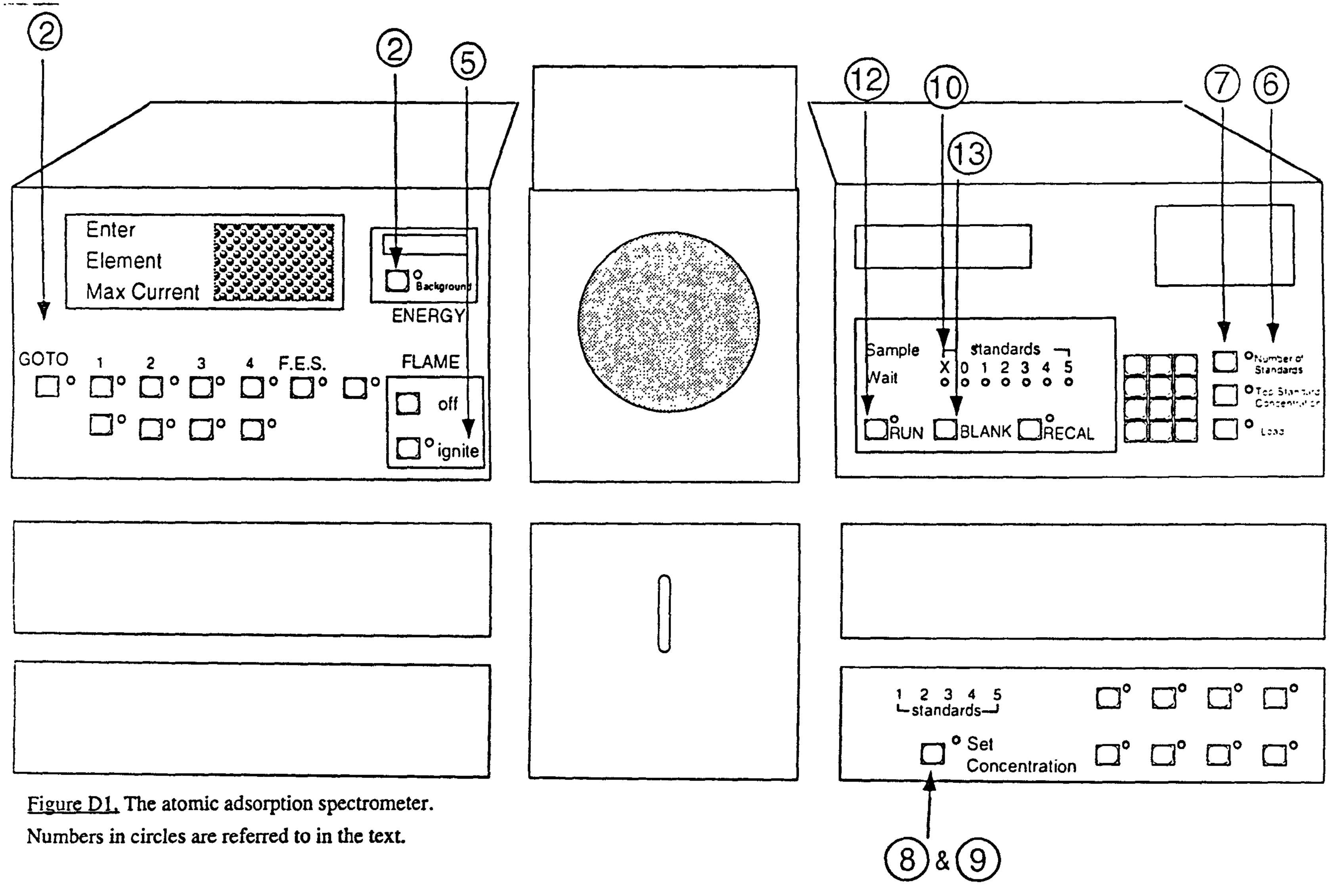


Figure D1. The atomic adsorption spectrometer.
Numbers in circles are referred to in the text.

analysis of aluminium meant that only one reading for the samples, could be taken for this element.

After all samples have been run, the machine is flushed with puite water to reamove any residual sample from the flame and then air is passed through the tubing for at least one minure, to enable the nebulizer and chamber to dry out. The flame is extinguished by pressing the 'off'button (5, on Fig. D1) and the acetylene and nitrous oxide cylinders and compressor are switched off. It is important to check at this point that the pressure gauge on each cylinder returns to zero. The machine is then turned off at the mains.

APPENDIX D4: Elements analysed and standards used

The AAS analyses elemental abundance based upon the weight percentage of the oxide of the particular element. Those elements studied are listed below, and the rock standards and particular AAS method employed for each given:

Sodium (Na_2O). Standards NIM-S, a syenite containing Na_20 by weight percentage; UB-N, a serpentine, 0.10wt.% Na_20; PCC-1, a peridotite, 0.01wt.% Na_20. Na_2O is analysed using flame emission spectrophotometry, using an air-acetylene flame.

Magnesium (MgO). Standard 400AMW, a dolomite containing 21.52 MgO by weight percentage; 401AMW, a limestone, 3.52wt. % MgO; GFS402(1), 5.63wt. % MgO.

Aluminium (Al_2O_3). AGV-1, an andesite containing 17.14Al_2O_3 by weight percentage; NIM-N, a norite, 16.50wt.% Al_2O_3; JB-2, a basalt, 14.67wt.% Al_2O_3. Al_2O_3 is determined by using a dinitrogen oxide-acetylene flame and atomic adsorption techniques.

Potassium (K_2O). Standards SCo, a shale containing 2.82 K_20 by weight percentage; SGR-1, a shale, 1.63wt.% K_20; BHVO-1, a basalt, 0.54wt.% K_20. K_2O is analysed using flame emission spectrophotometry, using an air-acetylene flame.

Calcium (CaO). Standard 400AMW, a dolomite containing 30.49 CaO by weight percentage; 401AMW, a limestone, 50.25wt. % CaO; GFS402(1), 46.82wt. % CaO. CaO is determined by using an air-acetylene flame and atomic adsorption techniques.

Manganese (MnO). Standards SY-2, a syenite containing 0.32 MnO by weight percentage; JB-1, a basalt, 0.15wt.% MnO; JG-1, a granite, 0.06wt.% MnO. MgO is determined by using an air-acetylene flame and atomic adsorption techniques.

Total iron ($FeO + Fe_2O_3$). Standard SCo-1, a shale containing 5.22 total iron by weight percentage; SGR-1, a shale, 2.98wt.% total iron; NIM-S, a syenite, 1.40wt.% total iron. Total iron is determined by using an air-acetylene flame and atomic adsorption techniques.

APPENDIX D5: Results

The results are summarised in sections 5.11-5.12 as percentage mass values for the oxides of each element. Each value represents the average of 3-4 separate runs of the sample through the machine (except for alumium**, see Appendix D3).

Appendix E

S.J. Metcalf, 1995

APPENDIX E1 : Whole rock grain size analysis by sieving

The grain size distribution of a sediment or soil unit is one of the basic descriptive features in sedimentology or soil science. In these analyses the grain size was expressed in terms of the scale proposed by J.A. Udden and C.K. Wentworth (Tucker, 1981), were sediments are divided into seven basic grades, which are (in order of increasing grain size) : clay; silt; sand; granules; pebbles; cobbles; and boulders (Tucker, 1981); as shown in Table E1. The sand and silt grades are further subdivided into five and four smaller classes respectively (Tucker, 1981) (Table E1). This scheme is based upon measurement of particle grain size in geometric SI units. However, the scale can also be converted into an arithmetic one by use of the logarithmic formula:

$$\phi = -\log_2 S \qquad \text{(C.1)}$$

Phi (ϕ) is the logarithmic transformation of the grain size (S) measured in millimetres (Tucker, 1981) and the resulting scale is shown in Table E1. The main advantage of using a phi scale instead of the geometric one is that it makes statistical calculations upon the data much simpler to do.

In the field, a rough approximation of the grain size for each soil horizon was made with a hand lens and comparison with the standard grain size scale card system. Following this, four samples were dried and weighed accurately in the laboratory (see below), before being passed through a stack of standard grain size sieves, ranging from -6phi units (>64μm) to 4phi units (<64μm). The four samples were originally taken from the clay lens prior to any excavation in 1987, by R.F. Vaughan (pers. comm., 1990) and were marked upon his preliminery log of the section as 'Beds B-E'. In terms of this thesis, they correspond to the following soil horizons:

Vaughan (1987)	This work	Sample weight
'Bed B'	Grey Clay 'BC'	210.52g
'Bed C'	Grey Clay '2Btg'	226.61g
'Bed D'	Green Clay 'Btg'	232.96g
'Bed E'	Green Clay 'Aae'	214.10g

Grain size units		Terminology		
mm	phi	Class terms	Subdivisions	Equivalent sedimentary terms
		boulders		
256	-8	----------		
128	-7	–		
64	-6	cobbles		gravel
32	-5	----------		rudite
16	-4	–		rudaceous sediments
8	-3			conglomerates
4	-2	pebbles		breccias
2	-1			
1	0	----------	very	
0.5	1	–	coarse	sand
0.25	2	granules	coarse	sandstones
0.125	3	----------	medium	arenaceous sediments
0.0625	4	–	fine	arenites
0.0312	5		very fine	
0.0156	6	sand	coarse	silt
0.0078	7		medium	siltstone
0.0039	8	----------	fine	
		–	very fine	clay
				claystone
		silt		

		–		
		clay		

Table E1. Grain size scale for sediments and sedimentary rocks (after Tucker, 1981)

In order to facilitate dry sieving of the sediments, the four samples were first passed through the finest grade sieve (<63μm or 4phi; Table E1) in order to remove the silt and clay. These factions tend to stick together in large lumps (or peds; see section 5.4) which are difficult to break apart, therefore these were wet sieved for a period of one-three hours, with occasional gentle agitation in order to fragment the larger lumps of clay. The residue was left to dry completely over a period of 48 hours and then passed through a stack of 12 dry sieves, by vigorously shaking the stack for up to 1/2 hour. The fractions remaining within each sieve was weighed in grammes and the percentage of sediment in each grade worked out for each sample. The results are shown in Table E2, along with the cumulative frequency expressed in percentages.

At this point the composition of the grains within each grade (above 63μm) was also ascertained. With the larger fractions (above 0.6mm), this involved separating by eye and weighing the different particles and converting them into percentage frequencies, whilst for the finer fractions a system of point counting particles under x10 and x40 magnification was used upon a small representative sample of the grade. These measurements were then converted to percentage frequencies and the results are shown in Table E3 (see also section 5.5).

The results are also represented graphically in section 5.3.3. Statistical tests were applied to the results and these are described and tabulated in section 5.3.3.

Grain size (phi)	Weight (g)	Frequency (%)	Cumulative (%)
'-6' - '-4'	79.25	37.64	37.64
'-4' - '-3'	4.93	2.34	39.98
'-3' - '-2'	8.05	3.82	43.80
'-2' - '-1'	2.60	1.24	45.04
'-1' - '0'	0.98	0.46	45.50
'0' - '0.75'	1.32	0.63	46.13
'0.75' - '1.5'	6.60	3.14	49.27
'1.5' - '2'	10.23	4.86	54.13
'2' - '3'	10.24	4.86	58.99
'3' - '3.5'	3.00	1.43	60.42
'3.5' - '3.75'	1.26	0.60	61.02
'3.75' - '4'	0.80	0.38	61.40
>4	81.26	38.60	100.00

(B)

Grain size (phi)	Weight (g)	Frequency (%)	Cumulative (%)
'-3' - '-2'	0.32	0.14	0.14
'-2' - '-1'	0.43	0.19	0.33
'-1' - '0'	0.45	0.20	0.53
'0' - '0.75'	0.83	0.37	0.90
'0.75' - '1.5'	6.21	2.74	3.64
'1.5' - '2'	13.15	5.80	9.44
'2' - '3'	15.62	6.89	16.33
'3' - '3.5'	4.71	2.08	18.41
'3.5' - '3.75'	1.74	0.77	19.18
'3.75' - '4'	1.00	0.44	19.62
>4	182.15	80.38	100.00

(C)

Grain size (phi)	Weight (g)	Frequency (%)	Cumulative (%)
'-1' - '0'	0.01	0.01	0.01
'0' - '0.75'	0.11	0.05	0.06
'0.75' - '1.5'	1.49	0.64	0.71
'1.5' - '2'	6.05	2.60	3.31
'2' - '3'	10.84	4.65	7.96
'3' - '3.5'	2.39	1.02	8.98
'3.5' - '3.75'	0.67	0.29	9.27
'3.75' - '4'	0.51	0.21	9.48
>4	210.90	90.53	100.00

(D)

Grain size (phi)	Weight (g)	Frequency (%)	Cumulative (%)
'-3' - '-2'	0.26	0.12	0.12
'-2' - '-1'	0.66	0.31	0.43
'-1' - '0'	1.87	0.87	1.30
'0' - '0.75'	3.29	1.54	2.84
'0.75' - '1.5'	8.02	3.74	6.58
'1.5' - '2'	9.71	4.54	11.12
'2' - '3'	15.72	7.34	18.46
'3' - '3.5'	4.00	1.87	20.33
'3.5' - '3.75'	1.08	0.50	20.83
'3.75' - '4'	1.16	0.54	21.37
>4	168.35	78.63	100.00

(E)

Table E2. Grain size data

Grain size (phi)	Calcareous	Ferruginous	Clay peds	Carbonaceous	Bone	Quartz
'-6' - '-4'	37.64	0	0	0	0	0
'-4' - '-3'	2.34	0	0	0	0	0
'-3' - '-2'	3.13	0.69	0	0	0	0
'-2' - '-1'	1.04	0.2	0	0	0	0
'-1' - '0'	0.4	0.05	0	0	0.01	0
'0' - '0.75'	0.53	0.07	0.01	0.01	0	0.01
'0.75' - '1.5'	2.63	0.13	0.13	0	0	0.25
'1.5' - '2'	1.5	0	0.23	0	0	3.12
'2' - '3'	1.08	0.13	0.3	0.12	0	3.23
'3' - '3.5'	0.22	0.03	0	0.06	0.03	1.09
'3.5' - '3.75'	0.13	0.06	0	0.03	0.01	0.37
'3.75' - '4'	0.13	0.04	0	0.01	0.01	0.19

(B)

Grain size (phi)	Calcareous	Ferruginous	Clay peds	Carbonaceous	Bone	Quartz
'-3' - '-2'	0.14	0	0	0	0	0
'-2' - '-1'	0.16	0.01	0	0	0.02	0
'-1' - '0'	0.14	0.05	0.01	0.01	0.01	0
'0' - '0.75'	0.22	0.03	0.09	0.01	0.01	0.03
'0.75' - '1.5'	1.19	0.15	0.32	0.06	0.11	3.61
'1.5' - '2'	1.67	0.26	0.26	0	0	3.61
'2' - '3'	0.71	0.14	0	0.07	0.07	5.9
'3' - '3.5'	0.09	0.04	0	0.04	0.05	1.86
'3.5' - '3.75'	0.02	0.01	0	0.01	0.01	0.72
'3.75' - '4'	0.02	0.01	0	0.01	0.01	0.39

(C)

Grain size (phi)	Calcareous	Ferruginous	Clay peds	Carbonaceous	Bone	Quartz
'-1' - '0'	0	0.01	0	0	0	0
'0' - '0.75'	0	0	0.01	0	0	0.04
'0.75' - '1.5'	0.01	0.09	0.07	0	0	0.47
'1.5' - '2'	0.04	0.13	0.11	0	0.01	2.31
'2' - '3'	0.09	0.32	0.17	0	0.02	4.05
'3' - '3.5'	0.02	0.05	0	0	0.03	0.92
'3.5' - '3.75'	0	0.02	0	0	0	0.27
'3.75' - '4'	0	0.02	0	0	0	0.19

(D)

(E)

Grain size (phi)	Calcareous	Ferruginous	Clay peds	Carbonaceous	Bone	Quartz	Oyster
'-3' - '-2'	0	0.12	0	0	0	0	0
'-2' - '-1'	0.09	0.11	0	0	0	0	0.11
'-1' - '0'	0.08	0.56	0.13	0	0	0.01	0.09
'0' - '0.75'	0.18	0.93	0.42	0	0	0.01	0
'0.75' - '1.5'	0.72	1.01	1.3	0	0	0.71	0
'1.5' - '2'	1.14	0.81	0.81	0	0	1.78	0
'2' - '3'	1.01	1.67	0.57	0	0	4.09	0
'3' - '3.5'	0.22	0.43	0.03	0	0	1.19	0
'3.5' - '3.75'	0.07	0.14	0	0	0.01	0.28	0
'3.75' - '4'	0.04	0.12	0	0	0.01	0.37	0

Table E3. Grain composition data

APPENDIX E2 : Fine fraction grain size anaylsis by sedigraph

An analysis of the silt and clay grades, i.e. particles below 63µm or 4phi, was carried out upon 34 samples taken from the soil blocks removed from excavation Site B. These samples correspond to those taken for petrological analysis and are shown in Fig. C2 (Appendix C). These samples were analysed using the Elzone machine, which is a sedigraph which counts grain size distribution based upon the electostatic conductivity of a solution made up of the individual sediment samples. In order to carry out the analysis, a 2g sample is put into solution with about 50ml of tap water The Elzone sedigraph was flushed though with electrolyte. Into a 100ml of electrolyte solution, 1ml of sample was pipetted and stirred vigourously, and this solution is termed the 'sample electrolyte'. The nozzle orifice on the Elzone was set to twice the estimated largest grain size of the sample (in this case 120µm) and the vacuum pump was turned on.

The sample was sucked through grain size nozzle and the sedigraph counted grain size by means of electrical conductivity of the sample. The Elzone machine produced a printout giving geometric mean grain size, the geometric standard deviation, the median, the mode, the arithmetic mean grain size and the arithmetic standard deviation. These measurements are expressed in SI units and as no statistical analyses were carried out upon the fine fraction alone, they were not converted into phi units. The average of three separate analyses are shown in Table E4. After each analysis, the machine was flushed out with pure electrolyte to normalize the counter.

Appendix F: Typical taphonomic data sheet (GLRCM G. 51201-G. 51264)

BOX	SP.	TYPE	GROUP	TAXA	REF.	CNTXT./DES	GRADE	GRID	MAX. LENGTH/cm	FEATURES	WEAR/BR(in life)	COMPLETE	FRACTURES	BREAKAGES
12	1	TO	CR:1		WH	sn. gry	A/B	W	0.45	RO(br)	N	Y	SP;LO	RO(sp)
12	2	TO	CR:1		WH	sn. gry	A/B	W	0.55	shed	TWF	Y	(SP),LO	N
12	3	TO	CR:1		WH	sn. gry	A/B	W	0.3	B(br)	N	B(br)	SP;(LO)	B(sp)
12	4	TO	CR:2		WH	sn. gry	A/B	W	0.3	shed	?TWF	T(br)	?TR;(LO)	T(tr-?wf)
12	5	TO	CR:1		WH	sn. gry	A/B	W	0.75	FRG-1/2	N	FRG-1/2	TR;SP	T(tr-split);B(sp)
12	6	TO	CR:1		WH	sn. gry	A/B	W	0.4	FRG<1/2	N	FRG<1/2	TR;SP;LO	T(tr-split);B(sp)
12	7	TO	CR:2		WH	sn. gry	A/B	W	0.4	NUB	N	T(br)	TR;SP;LO	T(sp/tr)
12	8	TO	CR:1		WH	sn. gry	A/B	W	0.45	shed	TWF	Y	SP;(LO)	T(sp-wf)
12	9	Plant	Wood	n/a	WH	sn. gry	A/B	W	0.5	FRG	n/a	FRG	n/a	flake
12	10	TO	CR:1		WH	sn. gry	A/B	W	0.3	2FRGS-1/2	TWF	2FRGS-1/2	TR;(LO)	tr-split
12	11	BO	indt	n/a	WH	sn. gry	A/B	W	0.2	SH-FRG	N	FRG	SP;LO	Y(sp)
12	12	TO	CR:2		WH	sn. gry	A/B	W	0.3	shed;FRG-1/2	N	FRG-1/2	TR;LO	tr-split
12	13	TO	CR:1		WH	sn. gry	A/B	W	0.3	shed	TWF	T(br)	TR;(LO)	T(tr-?wf)
12	14	BO	indt	n/a	WH	sn. gry	A/B	W	0.3	SH-FRG	n/a	FRG	TR;SP;LO	EP(sp/tr)
12	15	TO	CR:1	tiny	WH	sn. gry	A/B	W	0.2	RO(p)	N	Y	SP;(LO)	RO(sp)
12	16	TO	CR:1		WH	sn. gry	A/B	W	0.2	NUB	TWF-nub;scr	T(br)	SP;(LO)	T(sp-wf)
12	17	TO	CR:1	big	WH	sn. gry	A/B	W	0.8	RO(br);FRG-1/2	TWF	FRG-1/2	SP;TR;LO	RO(sp);tr-split
12	18	TO	CR:1		WH	sn. gry	A/B	W	0.45	shed	N	Y	SP;LO	RO(sp)
12	19	TO	CR:1	tiny	WH	sn. gry	A/B	W	0.25	shed	TWF-cxWF	T(br-wf)	(SP);LO	N
12	20	TO	CR:1		WH	sn. gry	A/B	W	0.35	shed	TWF	Y	LO	N
12	21	TO	CY	TR;ST?U	WH	sn. gry	A/B	W	0.35	U;CU(4)wf;VA(1);RO(br)	CUWF	FRG	SP;LO	FRG
12	22	TO	CY	TR;ST	WH	sn. gry	A/B	W	0.5	FRG;CU(2)wf;VA;RO(br)	CUWF(tip)	FRG->1/2	TR;SP;LO	FRG
12	23	TO	CY	TR;ST	WH	sn. gry	A/B	W	0.5	FRG;CU(2)wf;VA;RO(br)	CUWF(tip)	FRG->1/2	TR;LO	chipped
12	24	TO	CY	TR;ST	WH	sn. gry	A/B	W	0.4	FRG(w);CU(1);VA	N	FRG-<1/2	SP;LO	FRG
12	25	DA	CR		WH	sn. gry	A/B	W	?	keel;flat	n/a	FRG	SP;TR;(LO)	edges(sp/tr)
12	26	?SK	CR		WH	sn. gry	A/B	W	0.7	keel;curved	n/a	FRG	SP;LO	edges(sp)
12	27	TO	CR:1/2		WH	sn. gry	A/B	W	0.3	shed	TWF	Y	LO	N
12	28	TO	DI	TH;ind	WH	sn. gry	A/B	W	0.6	shed	TWF	Y	LO	N
12	29	?JA	PT		WH	sn. gry	A/B	W	0.6	FRG;SH?JA	n/a	FRG	SP;LO;(TR)	ends(sp);chipped
12	30	LB	FE	?DGT	WH	sn. gry	A/B	W	0.55	elongateLB;?DGT	n/a	br-EP(1)	SP;(TR;LO)	EP(sp),br(sp/tr)
12	31	LB	indt	n/a	WH	sn. gry	A/B	W	0.7	FRG;LB	n/a	br-EP(1)	SP;TR	EP(sp),br(tr)
12	32	TO	CR:1		WH	sn. gry	A/B	W	0 2	shed	N	T(br)	SP,TR	T(tr),RO(sp)
12	33	TO	CR 1		WH	sn gry	A/B	W	0 3	shed	TWF(+strip)	Y	(SP) LO	N
12	34	LB	LE	SP	WH	sn gry	A B	W	0 25	LB EP-large FRG	n a	br EP(1)	SP (TR) LO	EP broken chipped
12	35	?LB	HL(small)		WH	sn. gry	A/B	W	0.2	indtes	n a	br EP(1)	TR(sp)	EP broken chipped
12	36	TO	CR2		WH	sn. gry	A/B	W	0.35	shed	TWF	Y	LO	N
12	37	TO	PT	RH	WH	sn. gry	A/B	W	LOST	LOST	LOST	LOST	LOST	LOST
12	38	TO	PT/FI	RH	WH	sn. gry	A/B	W	0.25	conical;pointed	TWF;strip+//scr	FRG-T only	SP;TR	B(sp);chipped
12	39	TO	CR:3	large	WH	sn. gry	A/B	W	0.3	PMX;shed;L(ser)	TWF	NUB-T(br-wf)	SP;(LO)	T(sp-wf)
12	40	TO	CR:1		WH	sn. gry	A/B	W	0.35	shed;NUB	TWF;strip	NUB-T(br-wf)	SP;LO	T(sp-wf)
12	41	VE	?FI	TE?LP	WH	sn. gry	A/B	W	0.3	rnd;no prc;amph	n/a	FRG-1/2	(SP);TR	split(tr);abraded
12	42	TO	CR:1		WH	sn. gry	A/B	W	0.25	shed	TWF(tiny)	N	(LO)	N
12	43	TO	CY	TR;ST	WH	sn. gry	A/B	W	0.25	FRG;CU(1)wf;VA	CUWF	FRG<1/2	SP;TR;LO	FRG(sp/tr)
12	44	TO	CY	TR;ST	WH	sn. gry	A/B	W	0.15	FRG(tiny)	n/a	FRG<1/2	SP;LO	FRG
12	45	TO	CY	TR;ST	WH	sn. gry	A/B	W	0.25	FRG;CU(2);VA;RO(p)-br	N	FRG-1/2	SP;TR;LO	RO(sp);chipped
12	46	IN	GA	VV	WH	sn. gry	A/B	W		brown;replaced	n/a	N	n/a	N
12	47	IN	GA	NA	WH	sn. gry	A/B	W		FRG;cast	n/a	FRG	n/a	FRG
12	48	?LB	indt	n/a	WH	sn. gry	A/B	W	0.35	?LB;FRG-SH(noEP)	n/a	FRG-SH(no EP)	SP;TR;LO	EP/br(sp);chipped
12	49	?WP	?PT		WH	sn. gry	A/B	W	0.5	?WP;keel;flatten	n/a	FRG-SH(no EP)	SP;(TR);LO	EP/br(sp);chipped
12	50	SC	FI	TE;LP	WH	sn. gry	A/B	W	0.6	diamond	n/a	Y;edges br;flak	?SP	br(edges-sp)
12	51	SC	FI	TE;LP	WH	sn. gry	A/B	W	0.6	diamond	n/a	Y;edges br;flak	TR;(SP)	br(edges-tr(sp)
12	52	VE	indt	n/a	WH	sn. gry	A/B	W	0.25	VE(c);FRG;soggy bone	n/a	FRG<1/2;PR-br	SP;(TR)	br(sp);crumbled
12	53	LB	PT		WH	sn. gry	A/B	W	1.2	LBFRG-SH(noEP);?UL	n/a	FRG-SH(no EP)	SP;TR;LO	EP(sp);chipped
12	54	?LB	PT		WH	sn. gry	A/B	W	0.85	?LB;FRG(large)-SH	n/a	FRG-SH(no EP)	SP;LO;TR	EP(sp);chipped
12	55	SC	FI	TE	WH	sn. gry	A/B	W	0.2	FRG-flake	n/a	FRG	n/a	thin;smooth
12	56	SC	FI	TE	WH	sn. gry	A/B	W	0.25	DIA	n/a	Y;flaked	SP;LO	edges(sp)
12	57	SC	FI	TE	WH	sn. gry	A/B	W	0.6	IRG;edges eroded to point	n/a	IRG	SP;LO	edges-thin;smooth
12	58	DA/BO	indt	n/a	WH	sn. gry	A/B	W	0.3	FRG;indt	n/a	FRG	TR	edges(tr)
12	59	DA/BO	indt	n/a	WH	sn. gry	A/B	W	0.4	FRG;indt	n/a	FRG	TR;LO	edges(tr)
12	60	?SC	?FI	?TE	WH	sn. gry	A/B	W	0.35	IRG-FRG	n/a	FRG	TR;SP	edges(sp/tr)
12	61	SC	FI	TE	WH	sn. gry	A/B	W	0.5	IRG	n/a	IRG;flaked	SP;LO	edges(sp)
12	62	SC	FI	TE	WH	sn. gry	A/B	W	0.45	IRG	n/a	IRG;flaked	SP;LO	edges(sp)
12	63	SC	FI	TE	WH	sn. gry	A/B	W	0.6	DIA	n/a	Y;flaked	LO;(SP)	edges(sp)
12	64	SC	FI	TE	WH	sn. gry	A/B	W	0.45	IRG	n/a	IRG;flaked	SP;LO	edges(sp)

Table F1. Taphonomic data sheet

ABRASION	W. STAGE	SURFACE/EML	GM/BM	DIGESTION	TM	EROSIAN	CMPN	'TR' ABRASION	MATRIX	MINERALS
1	1	eml-worn patches=25%	N	N	N	N	N	n/a	N	N
1	2	strip(RB/C);>50%	N	?N	N	?Y	N	n/a	N	N
0(1)	0(1)	(crk)	N	N	N	N	N	n/a	N	N
0	0(1)	disc+pitted;<10%	N	?N	N	N	N	0	N	N
0	0	good	N	N	N	N	N	0	cly(cavity)	FeO(br surface)
2	2(3)	strip;30%;dentine-frosted	N	?Y	N	Y	?Y	0	?cly(br.surface)	?AP(br surface)
1	3	strip;60%;dentine-frosted	N	Y	N	Y	?Y	0(1)	?cly(strip)	?AP(strip)
0(1)	1	strip(B)<10%	N	?N	N	N	N	n/a	N	N
n/a	n/a	n/a	n/a	n/a	n/a	Y	?	n/a	N	FeO(rep)
0	1	strip;pitted;>60%	N	?N	N	N	N	0	?lst(strip)	N
1(2)	1(2)	crk	N	n/a	?5%	?ww	N	n/a	N	N
0(1)	0(1)	(crk)	N	N	N	N	N	0(1)	N	N
0	0(1)	(crk)	N	N	N	N	N	1	white	N
1(2)	2	crk	N	n/a	N	?ww	?N	0(1)	Y(cavity)	N
0	0(1)	(crk)	N	N	N	N	N	n/a	N	N
0(1)	0	good	N	N	N	N	N	n/a	N	N
0(1)	2	strip;10%+pitted	N	?N	N	?N	?N	0	N	?FeO(in crks)
0	1(2)	crk;pitted+strip(RB);<10%	N	?N	N	?N	N	n/a	white(surface)	N
1(2)	1(2)	strip-wf onto cnvx;20%	N	N	N	?N	N	n/a	cly(surface)	N
2(3)	2(3)	strip(RB);30%,frosted	N	?N	N	Y	N	n/a	N	N
2(3)	3	crk;strip CU-?wf	N	N	N	?Y	N	n/a	ooid+cly	N
1	2	crk	N	N	N	?Y	?N	0(1)	lst+cly(br)	N
1	2	crk	N	N	N	?N	?N	0	cly(base)	N
1(2)	1(2)	crk	N	N	N	?Y	?N	n/a	N	N
2	1	(crk)	N	N	N	?ww	N	2	cly(surface pits)	N
3	2	abraded	N	N	N	?ww	N	n/a	N	N
0	1(2)	crk	N	N	N	N	N	n/a	N	N
2	2(3)	strip-(lbl)>50%	N	?N	?10%	?N	N	n/a	N	?FeO(in crks)
1	1(2)	crk	?BM	N	?5%	N	?N	0		
1(2)	1	EPstrip(abraded-bbly)	N	N	N	?ww	N	1	N	N
1(2)	0(1)	good	N	N	N	N	?Y	1(2)	N	N
0	0	good	N	N	N	?N	N	0	grn cly+ooid	N
1	3	strip(from TWF)50%;frosted	N	?Y	N	?Y	N	n/a	cly	N
1(2)	2	crk	N	N	N	?N	?N	0	cly(surface)	N
1(2)	3	strip/abraded	N	?N	N	ww	Y	1	cly	N
0(1)	2	sanded(RG-bsl);35%	N	?Y	N	N	?N	n/a	N	N
LOST	LOST	LOST	LOST	LOST	LOST	LOST	LOST	LOST	LOST	LOST
0(1)	2	strip(faces)+scr	N	N	<5%	Y	N	0	grn cly(bsl cavty)	?N
0	0(1)	fair	N	N	N	N	N	n/a	N	N
2	3	strip(from TWF)	N	N	N	N	N	n/a	N	N
1(2)	3	surface lost 100%	N	?N	N	Y	Y	1	cly(in pores)	?AP/CLT(in pores)
0	1	(crk);pitted 5%	N	N	N	N	N	n/a	N	N
0(1)	2	crk	N	N	N	?Y	Y	0	N	FeO(surface)
2(3)	2(3)	crk	N	?N	N	?Y	?N	n/a	cly(surface)	?grn(surface
0(1)	1(2)	crk	N	N	N	?N	?N	0(1)	cly+ooid(RO)	N
0	0	F(detail visible)	n/a	N	n/a	N	N	N	N	N
2(3)	3	cast(abraded)	n/a	N	n/a	Y	Y	n/a	N	CaCO3 replaced
1	1(2)	fair;crk	N	N	N	N	N	1	grn/gry cly	N
1	3	crk;strip	N	?N	N	?Y	N	1	N	N
1(2)	1(2)	flaked eml;pitted	N	?N	N	?N	N	n/a	N	N
2	3	flaked eml	N	?N	N	?N	Y	0	N	N
1	2(3)	edges abraded-bbly;crumbly	N	N	N	?N	?	?	gry cly-consolidant	N
1	2(3)	striated-abraded+encrusted	N	?N	N	?Y	N	0	grn cly+ooid(surface/cavity)	FeO(surface)
1	2	striated;split	N	N	N	Y	Y	0	N	N
0	3	flake of shiny eml	N	N	N	N	N	n/a	N	N
2(3)	3	flaked	N	N	N	?Y	N	n/a	N	N
2(3)	3	smooth;not flaked	N	Y	N	Y	N	n/a	cly(surface)	N
n/a	2(3)	N	N	?Y	N	Y	?	1	cly(surface)	N
1	2	pitted	N	?Y	N	Y	?		N	N
2	0(1)	N	N	N	N	?N	N	0	cly(surface)	N
3	3	flaked(smooth)	N	?Y	N	Y	N	n/a	cly(surface)	N
2(3)	3	flaked(smooth)	N	?Y	N	Y	N	n/a	cly(surface)	N
0(1)	2(3)	flakey	N	N	N	Y	N	n/a	N	N
2	3	flaked(smooth)	N	?Y	N	Y	N	n/a	cly(surface)	N

Table F1. continued....